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THE COL-GULLY AND GLACIAL DEPOSITS AT COURT HILL, CLEVEDON, NEAR BRISTOL, ENGLAND

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ABSTRACT. An outline is given of the Quaternary geology and geomorphology of Court Hill Col in Failand Ridge near Clevedon, Avon County, from observations made during the construction of the M5 Motorway.

A glacial col-gully about 100 m wide and approximately 25 m deep is described. The col-gully, eroded through the Carboniferous Limestone, opens and deepens northward. Associated with the Col and the col-gully is a complex sequence of Quaternary deposits. Uppermost in the sequence is a layer of red sandy silt (cover sand) approximately 0.5 m thick, of periglacial origin, probably of Devensian (Weichselian) age. Largely confined to the col-gully are unstratified tills, stratified ice-contact deposits and glacio-lacustrine deltaic deposits. The glaciogenic deposits are up to 25 m thick. Boulders of about 8 Mg in weight have been observed.

The geomorphology of the col-gully, and the stratification and composition of the glaciogenic deposits, demonstrate that an ice sheet at least 85 m thick had impinged against the south flank of Failand Ridge and was discharging immense quantities of water and sediment down an ice-contact slope through the Col into a small ice-marginal lake north of the col-gully. The ice sheet is regarded as being Wolstonian, or Anglian, in age.

The precise origins of the col-gully and the interpretation of the glacial sequence are not yet completely clear. However, it is believed that the balance of evidence indicates that both the col-gully itself and the glaciogenic deposits represent a complex sub-, en- and pro-glacial sequence associated with the downwasting and division of an ice mass into two parts by the "emergence" of Failand Ridge. The possible extent and geomorphological implications of ice-sheet penetration into the Bristol area are briefly discussed.

RÉSUMÉ. *Les dépôts glaciaires et torrentiels de Court Hill, Clevedon, près de Bristol, Angleterre.* On donne un aperçu de la géologie et de la géomorphologie du quaternaire du Court Hill Col dans la Failand Ridge près de Clevedon, Avon County, à partir d'observations réalisées durant la construction de l'autoroute M5.

Un défilé glaciaire d'environ 100 m de large et approximativement 25 m de profondeur est décrit. Le défilé, creusé dans les calcaires du carbonifère, s'ouvre et s'approfondit vers le nord. Associée avec le Col et le défilé, on trouve une séquence complexe de dépôts quaternaires. Au sommet de la séquence il y a un niveau de sables rouges limoneux (sables de couverture) épais d'environ 0,5 m, d'origine périglaciaire d'âge probablement Devensian (Weichselian). Largement limitées au défilé sont des argiles morainiques non stratifiées, des dépôts stratifiés de contact à la glace et des dépôts déltaïques glacio-lacustres. Les dépôts d'origine glaciaire ont jusqu'à 25 m d'épaisseur. Des blocs d'environ 8 Mg en poids ont été observés.

La géomorphologie du défilé, la stratification et la composition des dépôts d'origine glaciaire démontrent qu'une calotte glaciaire d'au moins 85 m d'épaisseur s'est bloquée contre le flanc sud de la Failand Ridge et débitait d'énormes quantités d'eau et de sédiments au bas d'une pente au contact avec la glace à travers le Col dans un petit lac pro-glaciaire au nord du défilé. On pense que la calotte est d'âge Wolstonian ou Anglien.

Les origines précises du défilé et l'interprétation de la séquence glaciaire ne sont pas encore complètement claires. Cependant, on pense que le bilan des preuves indique que tant le défilé lui-même que les dépôts d'origine glaciaire représentent une séquence complexe sub-, en- et pro-glaciaire associée à la destruction et à la division d'une masse de glace en deux tronçons par l'"émergence" de la Failand Ridge. On discute brièvement de l'extension possible et des implications géomorphologiques d'une pénétration d'une calotte glaciaire dans le secteur de Bristol.

ZUSAMMENFASSUNG. *Die rinnengebundenen und glazialen Ablagerungen am Court Hill, Clevedon nahe Bristol in England.* Die Quartär-Geologie und -Geomorphologie des Court Hill Col im Failand Ridge bei Clevedon, Avon County, wird auf der Grundlage von Beobachtungen beim Bau der Autobahn M5 dargelegt.

Eine glaziale Joch-Abflussrinne von etwa 100 m Breite und annähernd 25 m Tiefe wird beschrieben. Die Rinne führt durch Karbonischen Kalkstein und verläuft unter Eintiefung nordwärts. Verbunden mit dem Joch und der Rinne ist eine komplizierte Folge quartärer Ablagerungen. Zuoberst in der Folge findet sich eine etwa 0,5 m dicke Schicht aus rotem, verschlammtem Sand (Decksand) periglazialen Ursprungs, vermutlich aus der Devensian- (Weichsel-) Eiszeit. Im wesentlichen auf die Rinne beschränkt, folgen ungeschichtete Schotter, geschichtete Eisrand-Ablagerungen und deltaartige Aufschüttungen in einen Eisrandsee. Die glaziogenen Ablagerungen sind bis zu 25 m mächtig; sie enthalten Felsbrocken bis zu 8 Tonnen Gewicht.

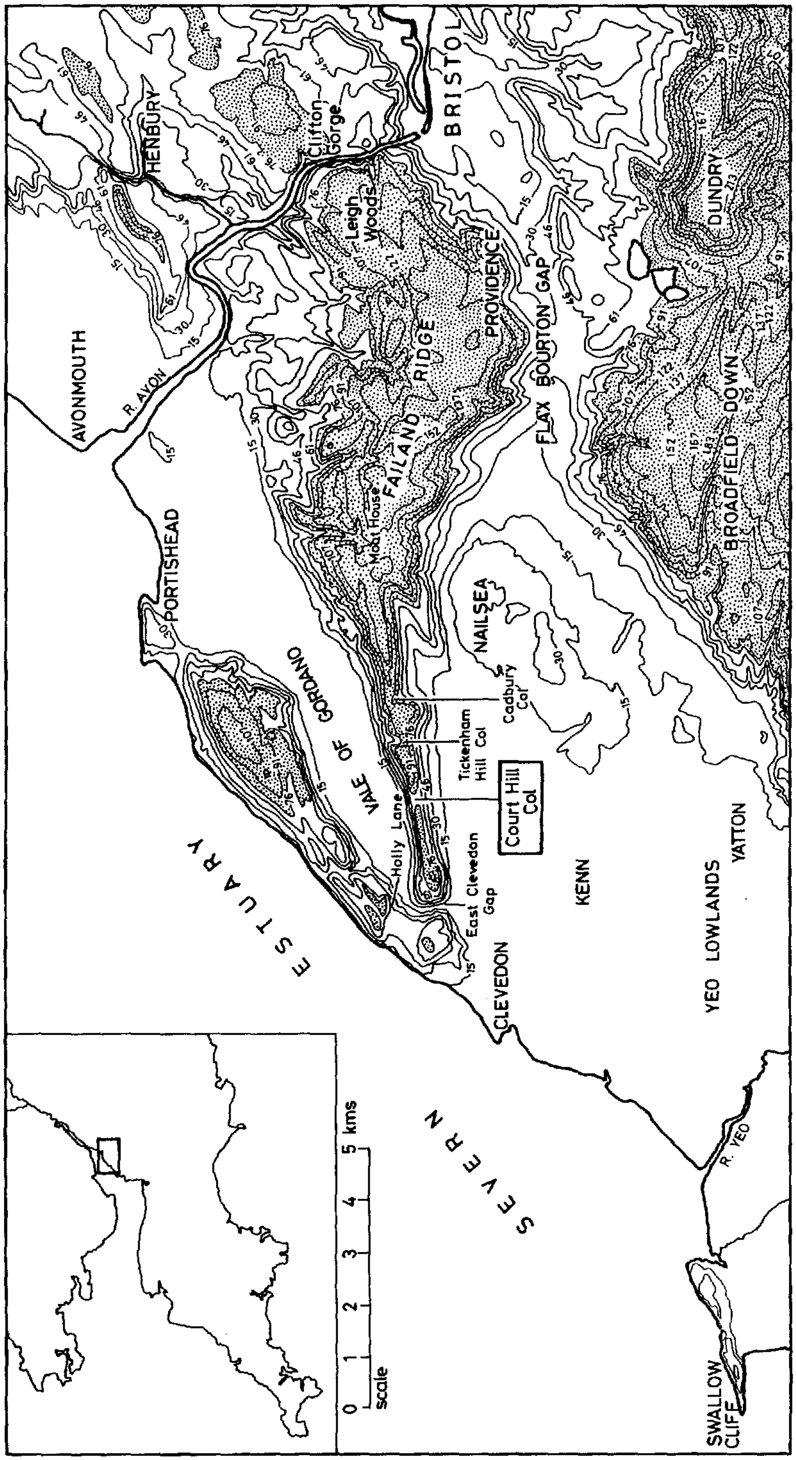

Fig. 1. Topographic map to indicate the position and relationship of Court Hill Col.

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Die Geomorphologie der Abflussrinne und die Schichtung und Zusammensetzung der glaziogenen Ablagerungen lassen erkennen, dass eine Eisdecke von mindestens 85 m Dicke sich an die Südflanke des Failand Ridge herangeschoben hatte und riesige Mengen von Wasser und Schutt über einen Hang am Eisrand durch das Joch in einen kleinen Eisrandsee nördlich der Jochrinne ergoss. Die Eisdecke dürfte der Wolstonian- oder Anglian-Eiszeit angehört haben.

Der Ursprung der Jochrinne und die Deutung der glazialen Ablagerungsfolge bedürfen noch weiterer Untersuchungen. Doch lässt die Gesamtheit der Erscheinungen glaubhaft erkennen, dass sowohl die Jochrinne selbst wie die glaziogenen Ablagerungen ein kompliziertes sub-, in- und proglaziales System darstellen, das mit dem Verfall und der Spaltung einer Eismasse in zwei Teile durch den "Aufstieg" des Failand Ridge verbunden war. Das mögliche Ausmass und die geomorphologischen Folgen des Vordringens der Eisdecke in das Gebiet von Bristol werden kurz diskutiert.

Introduction

Failand Ridge forms an area of high ground stretching from Clifton Gorge at Bristol in the east to the Severn Estuary at Clevedon (Fig. 1). Principally, it is composed of Carboniferous Limestone. Three main cols occur in the western part of the ridge—Court Hill Col, Tickenham Hill Col and Cadbury Col. In addition, the ridge is completely breached by the steep-sided, deep through valley of East Clevedon Gap (also known as "Swiss Valley").

In 1853, Joshua Trimmer described the drift deposits he located in Norton's Wood (Fig. 2), probably at grid reference ST 437723. Trimmer concluded from the content of "non-local" rock types and the altitude of the deposits that they may represent a southern termination of the "glacial sea", then associated with what we now know as glaciation. Between 1970 and 1972, the drift observed by Trimmer was exposed during the construction of the M5 Motorway through the Col. The then available exposures at Court Hill, and other sites in the area, were demonstrated and described briefly by Hawkins and Kellaway (1971). These observations together with subsequent studies have led several authors to re-advance Harmer's (1907) much criticized hypothesis that Pleistocene ice sheets penetrated far into the coastal lowlands of what was then Somerset (since 1974 partly in Avon County). The recent finds of glaciogenic deposits plus the current interpretations of geomorphological evidence do not necessarily agree in detail with those advanced by Harmer (Donovan, 1969; Hawkins and Kellaway, 1971; Hawkins, 1972; Colbourne and others, 1974; Gilbertson and Hawkins, in press).

The present paper describes more fully and interprets the Quaternary deposits and landforms exposed in Court Hill Col, as observed during the construction of the M5 Motorway (Fig. 3). The Col is significant for several reasons. First, it demonstrates the penetration of an ice sheet into a part of southern England which until 1970 had been considered by most

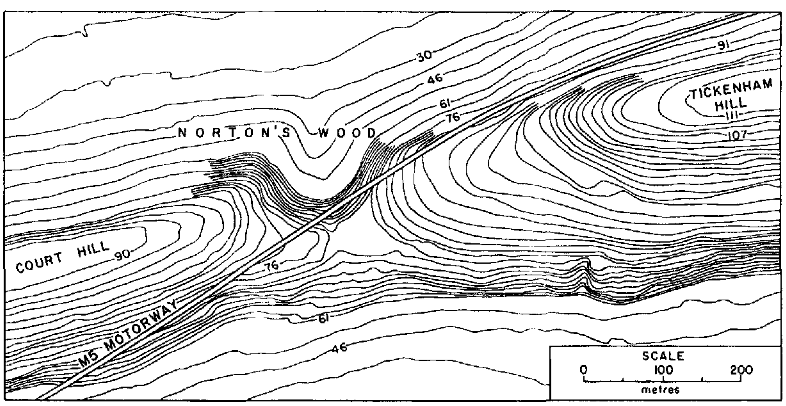

Fig. 2. Topography of Court Hill Col prior to the beginning of construction for the M5 Motorway. Contour interval changes above 61 m O.D.

authors to be beyond the limit of glaciation. Secondly, it is an example in lowland Britain of the glacial landform originally described from Scandinavia by Mannerfelt (1945) as a sadelskar or col-gully. Further, the motorway excavations revealed in detail the internal structure of the glaciogenic deposits occupying the col-gully.

Fig. 3. Topographic setting of Court Hill Col.

As a result of slope grading and seeding, many details of the exposures studied during construction are now lost. It is, however, still possible to determine the col-gully walls from the bridle-path bridge over the motorway. Similarly, blocks of naturally cemented drift are still visible by the motorway boundary fence at Norton's Wood, and by the eastern foundations of the bridle-path bridge. Access to the main motorway cutting slopes is prohibited.

The stratigraphical nomenclature used here follows that recommended for description of the British Quaternary in Mitchell and others (1973).

Relief and geology

The surface relief of Court Hill Col, prior to the motorway construction, is shown in Fig. 2. The Col itself was 74 m O.D. and approximately 0.25 km wide, lying between Court Hill (90 m O.D.) to the west and Tickenham Hill (111 m O.D.) on the east. Recent reference to the geology of this area was made in Hawkins and Kellaway (1971) and Hawkins (1972). The Col is eroded through the Black Rock Limestone and Dolomite of the Carboniferous Limestone Formation, dipping at about 35° south. Here, the Carboniferous limestones are thrust over Coal Measures sandstones, which crop out on the north side of the Col. Dolomitic Conglomerate and Keuper Marl of the Triassic crop out south of the crest of the ridge. Both flanks are covered by a varying thickness of red sandy silt, initially mapped by Osmond as the Tickenham Soil Series and subsequently shown to be a cover sand of periglacial origin (Findlay, 1965; Gilbertson and Hawkins, in press; Vink, unpublished).

The area of Court Hill Col was extensively investigated by drilling prior to the construction of the M5 Motorway. Particle-size analyses by Soil Mechanics Ltd (in 1967) revealed the presence of well to poorly sorted sand, gravel, and cobble layers overlying the Black Rock Dolomite. The distribution of this drift was mapped by Hawkins and Kellaway (1971, fig. 5). Their results, together with subsequent examination, have shown that the drift crops out from about 83 to 65 m O.D. on the south flank of the Col, while the lowest outcrop on the north flank, in Norton's Wood, is about 50 m O.D. The total east–west width of the drift, beneath the cover sands, is approximately 130 m. The thickness of the drift beneath the cover sand varies, the great bulk of the drift being confined to the deep channel or gully—the col-gully—eroded into the Carboniferous Limestone bedrock. Up-slope of the main col-gully, the cobble- and boulder-rich deposits thin very rapidly.

Description

Stratigraphy of Pleistocene deposits

The thickness and stratigraphical position of the various deposits associated with the Col have been reconstructed from many temporary exposures. A composite section corresponding to the line of the central motorway reservation is illustrated in Figure 4 and described below.

Unit 6. Red sandy silt. This uppermost deposit covers both drift and bedrock to a depth up to 0.5 m. These soils can be traced into the Tickenham Soil Series of Findlay (1965). Open solution joints in the Carboniferous Limestone are frequently infilled with sandy silt for a depth of several metres (probably re-deposited cover sand).

Unit 5. Red-brown gravel to boulder-rich diamicton. The deposit is unsorted and unbedded. Boulders with diameters up to 0.5 m are common. The matrix is a stiff red sandy silt. The thickness is variable but greatest near the eastern wall of the col-gully, where it reaches up to 3 m.

Unit 4. Diamicton up to 0.75 m thick; comprising cobbles and boulders in a poorly bedded sandy silt matrix. This layer may be impersistent; the excavations did not permit this to be checked.

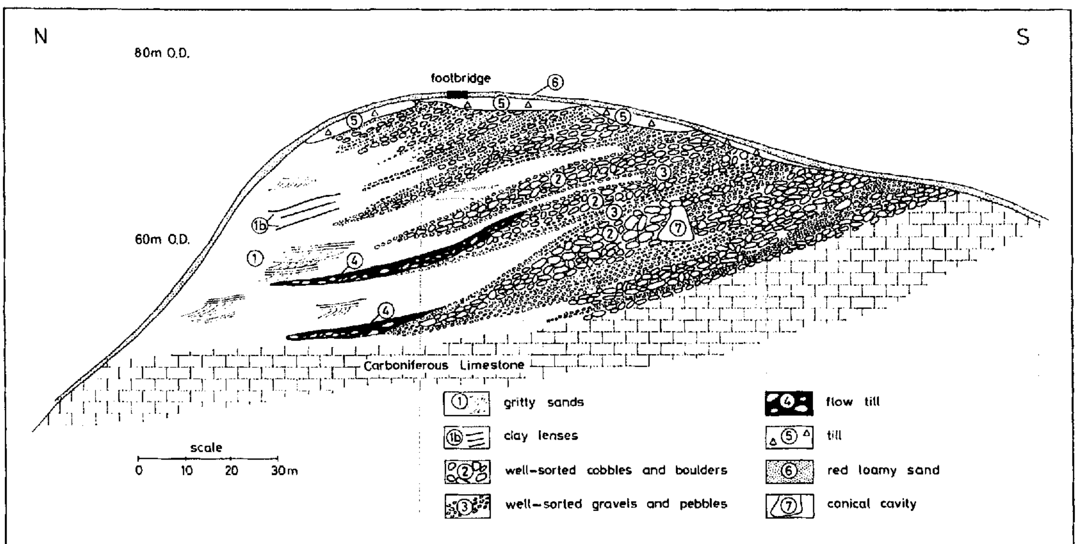

Fig. 4. Schematic section through the Quaternary deposits in Court Hill Col.

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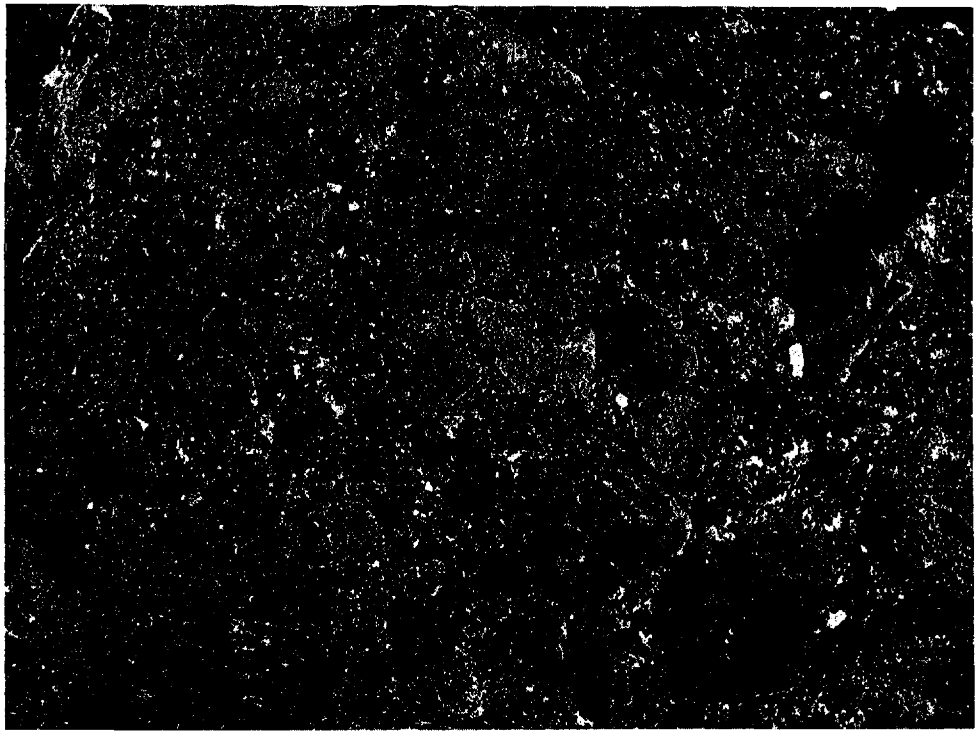

Fig. 5. Partially cemented (point contact) interbedded gravels with cobbles and boulders. The beds dip northward up to 34°.

Fig. 6. Well-sorted gravels giving a more compact material.

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Unit 3. Gravel and cobble layers between 0.5 and 1.5 m thick; well-sorted, generally well-rounded, sometimes bedded and often grading smoothly upwards or downwards into unit 2. Calcareous cementation at point contact. In some cases more densely packed (Fig. 5).

Unit 2. Stratified gravel to boulder deposit, with rare silty sand matrix in the finer material. Generally well-sorted well-rounded clasts but in some horizons the clasts were less well sorted and rounded. In places calcareous cement binds the material; however, where there is little or no matrix, the clasts are cemented at point contact. Boulders up to 8 Mg in weight have been observed. Large conical cavities occur measuring up to 2–3 m deep, 1 m wide at the top and 3–4 m at the base (unit 7). In the northern half of the section, the deposits could be seen to form 0.2–2 m bands of material dipping northward at 30–37° (Fig. 6) through the Col, not parallel to the dip of the col-gully walls.

The deposits of unit 2 grade into those of units 3 and interdigitate with material of unit 1.

Unit 1. Coarse gritty sands interdigitating with the boulder-rich deposits in the northern half of the col-gully. The sands are markedly cross-laminated (Fig. 7). Occasionally, the sands are interspersed with thin clay/silt seams, always less than 20 mm in thickness (unit 1b).

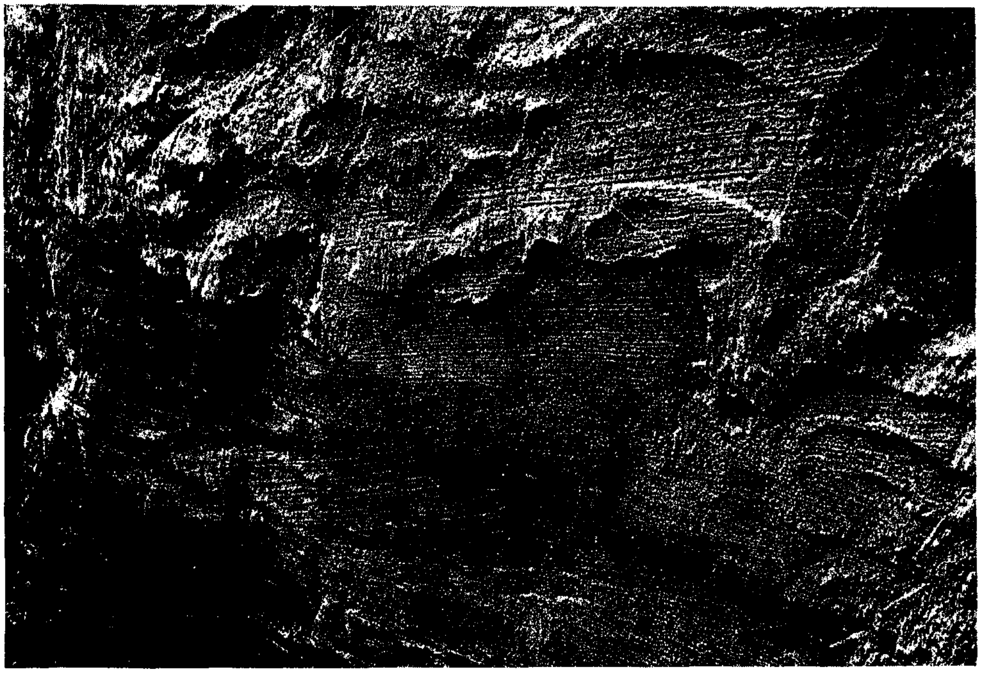

Fig. 7. Cross-bedded loose sands, as they occur "below" the main gravelly layer.

Composition

The gravel, cobble and boulder deposits are cemented at point contact by a calcareous deposit. Apart from unit 6, the composition of the main deposits is similar. The largest clasts and the most frequent rock type are Carboniferous Limestone. Other fairly common rock types were Pennant Sandstone (Carboniferous), chert (Carboniferous), yellow Triassic sandrock and irregular "lumps" of Keuper Marl (Triassic), and red Devonian sandstones. Also present are smaller quantities of Cretaceous Chalk flints and Cretaceous Greensand cherts.

The sands of unit 1 yielded the following relic Jurassic and Cretaceous foraminiferal fauna, identified by Professor J. W. Murray: *Marginulina, Frondicularia, Saracenaria, Citharina, Nodosaria, Hedbergella, Praeglobotruncana, Gyroidina*, and *Gumbitriella.*

The col-gully

The detailed location and dimensions of the sub-drift col-gully are imperfectly known from direct observation and bore holes. Figure 8 shows the geology of the gully as determined from bore holes. As frequently with engineering schemes, however, the bore-hole information is not adequate for the purpose of this study and does not provide a complete morphological section across the gully. Further drifts often obscure the surface outcrop. Where observed, both col-gully walls had a slope of more than 30° from the horizontal.

The col-gully opens and deepens northward, its floor falling from *c.* 65 m to *c.* 50 m O.D. (Fig. 4). It opens northward, the east wall being orientated at 25° N., the west wall at 355° N. Morphologically, the col-gully "hangs" above the adjacent alluvial lowlands, which lie at about 7 m O.D.

INTERPRETATIONS

Two general interpretations may be advanced: (a) Triassic origin and (b) Pleistocene origin.

Triassic interpretations

Triassic and conglomeratic deposits (Dolomitic Conglomerate) filling hollows and gullies in the Carboniferous Limestone are well known in the Bristol district (Kellaway and Welch, 1948). The Court Hill col-gully and its associated deposits, however, cannot be the result of Permo-Triassic processes as Cretaceous rock types and Cretaceous and Jurassic fossils are present in the deposits.

The East Clevedon Gap is shown on Geological Sheet 264 to be floored by Keuper Marl (Triassic). Hawkins (1972), however, reported that a bore hole in the middle of the valley failed to prove the presence of Keuper Marl below the post-Cretaceous gravels and sands, which he interpreted as being Quaternary.

Pleistocene interpretations

The drift deposits. There is considerable evidence to suggest the col-gully and its infill deposits are Pleistocene in age.

Unit 6. The stratigraphical relationship of the red sandy silt (loam) covering all lower drifts and bedrock on Failand Ridge suggests it is aeolian in origin. This conclusion is supported by mineralogical and particle-size investigations, which have shown it is an exposure of part of the almost continuous blanket of periglacial cover sand which mantles many of the Carboniferous Limestone hills south of Bristol (Findlay, 1965; Colbourne and others, 1974; Gilbertson and Hawkins, 1974, in press).

Unit 5. The following lines of evidence indicate this cobble- and boulder-rich diamicton is a till and *not* a soliflucted Triassic deposit into which cobbles and boulders of local and non-local provenance have been incorporated by frost processes. It is unsorted, unbedded and has a stiff red sandy silt matrix binding together substantial boulders (many about 0.5 m in diameter) and cobbles of Devonian, Carboniferous, and Cretaceous rock types distributed fairly uniformly through the deposit. Although this diamicton is dependent for its red fine fraction upon incorporation of Keuper Marl, it occurs above the height of *in situ* or soliflucted Triassic deposits on this part of Failand Ridge. *In situ* Devonian, Jurassic, and Cretaceous rock types do not occur in the immediate area.

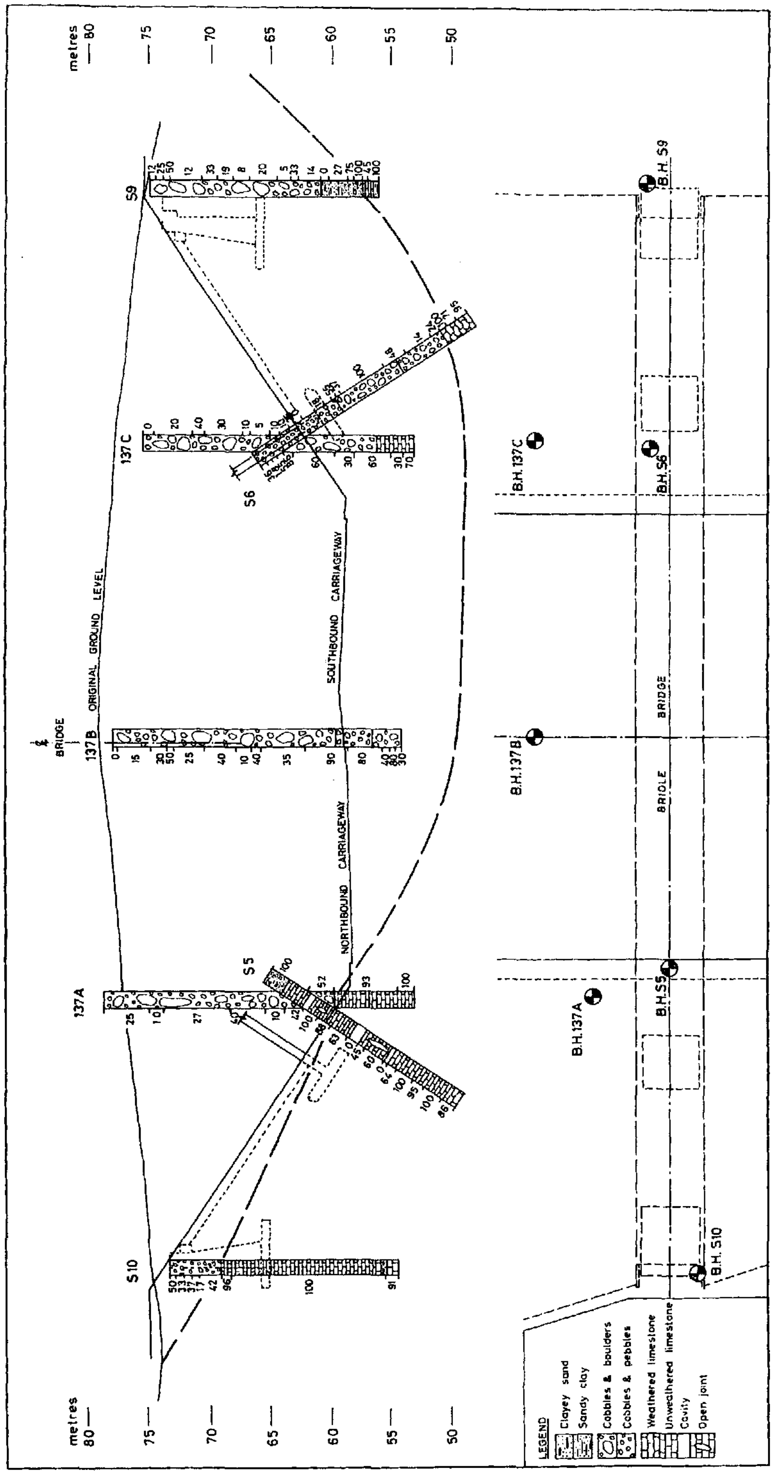

Fig. 8. Bore-hole data and suggested profile through the Quaternary valley (col-gully) based partly on field evidence in other areas of the valley. Figures against the bore-hole logs give percentage core recovery.

Unit 4. The stratigraphical relationships of the layers of clay/silt with cobbles and boulders, overlying or replacing the gravel, cobble, or boulder bands of units 3–2 suggest these drifts have slid or slumped down the developing drift surfaces and, as such, have the characteristics of flow tills.

Units 3–2. The steep (30–37°) dip, rapid changes in clast size (gravel to boulders of up to 8 Mg), good to poor sorting and rounding are, taken together, properties characteristic of ice-contact deposits (Flint, 1971). In the topographic context of the Court Hill col-gully, the steeply dipping bands of coarse and fine material generally washed free of fines (units 3–2) can only be satisfactorily explained as the result of large fluctuations in discharge down a steep ice-marginal slope. Seasonal or even diurnal variations in temperature may be indicated.

The large cavities (unit 7) present in the most boulder-rich deposits in the south of the gully only seem explicable if considered as voids left by the melting of ice blocks incorporated in the sediment; in this case the cementing of the boulders at point contact by a calcitic cement must have taken place rapidly, since few signs of wall and roof collapse have been seen in the voids inspected. This cement may have been derived from the solution of calcareous rocks—Carboniferous, Jurassic, or Cretaceous—or perhaps from shells originally derived from the floor of the Bristol Channel or Irish Sea area.

Unit 1. These gritty cross-laminated sands with topsets, foresets, and bottomsets are obviously waterlain. Their interdigitation with ice-contact sediments suggests they are ice-marginal in origin and may be referred to as "glacio-lacustrine". There is no indication of wave or current activity in the sedimentary structures. The thin silt bands indicate that on occasion the water body became sufficiently quiet for fines to settle out on the bottom. A water level of at least 75 m O.D. occurred at one period. The dimensions of the sand body indicate a small lake at least 50 m by 150 m in size must have been present. The physiography (Figs 1 and 2) of the area indicates that a lake or any other body of water can only have occurred in one of two ways: either the Vale of Gordano (Fig. 1) was occupied by an ice mass, in which case the water body had a surface area measurable in thousands of square metres; or alternatively, if the Vale of Gordano was ice-free, the normal drainage of the area must have been obstructed by an ice mass stretching across and blocking drainage in the Severn Estuary. This would have produced pro-glacial Lake Maw (Mitchell, 1960, 1972), measurable in hundreds of square kilometres.

Composition. The precise provenance of the rock types remains uncertain since it is possible to conceive that the former high-level drainage networks, drainage captures and subsequent re-cycling patterns discussed by Frey (1975) could all bring the various rock types and fossils to the area. Nevertheless, there are no known remanié deposits in the area which would yield either the volume of material or the individual clast sizes found at Court Hill.

Four rock types merit particular attention. Coal Measures sandstone is conspicuously lacking in the drift. This is surprising in view of its outcrop immediately north of the Col and suggests the major area of derivation was to the south of the line of Failand Ridge. Greensand cherts are common in most drifts of the west of England but Cretaceous flints are much less common. The cherts characterize the drifts of the Bristol district, although the area is well separated from the Greensand scarp. Unlike the siliceous cherts and flints, the calcareous Cretaceous micro-fossils seem unlikely to resist multiple re-cycling from higher surfaces and drainage networks. Consequently, there appear good grounds for regarding Greensand chert, Chalk flints and Cretaceous micro-fossils in the Bristol area as, at least in part, resulting from an introduction by ice sheets. Recent investigations of the submarine geology of the sea bed west of Bristol (Donovan and others, 1961, 1970; Curry and others, 1967; Hamilton and Blundell, 1970; Lloyd and others, 1973) lend support to the hypothesis advanced in Gilbertson and Hawkins (in press) that, as an ice sheet moved into the Somerset lowlands from the Bristol

Channel and Celtic Sea area, it would have incorporated all the rock types found at Kenn and now at Court Hill.

The col-gully and glacial sequence at Court Hill. In view of the topographic and geological setting of the Court Hill Col discussed previously, it is difficult to conceive of any process, other than glacial, which would provide a sufficient volume of water to erode the col-gully at the appropriate altitude. This observation is supported by the morphological and dimensional similarities between the Court Hill col-gully and the col-gullies originally described from Scandinavia by Mannerfelt (1945) and from Britain by Derbyshire (1961) and Clapperton (1968). These landforms have been identified with erosion by glacial melt waters. The extent to which the northern exit of the Court Hill col-gully "hangs" above the Vale of Gordano suggests the vale was occupied by water, and/or ice, to a height of approximately 65 m O.D., at the same time as the col-gully was being eroded. If this were not the case, it would be anticipated that the gradient of the col-gully floor would merge with the form of the hillslope or floor of the Vale. Irrefutable evidence for the occupation of the vale by ice sheets as opposed to water is as yet lacking but several authors have suggested that its Pleistocene drifts with erratics, its surprisingly large buried channel system and the anomalously routed valleys on its hillslopes may well be explained as the result of the former downwasting ice mass (Hawkins, [1968], 1972; Hawkins and Kellaway, 1971; Colbourne and others, 1974; Gilbertson and Hawkins, 1974, in press). Occupation of the Vale of Gordano by ice would also explain the water body in which the gritty sands (unit 1) collected as a comparatively small ice-marginal lake trapped against the north flank of Failand Ridge rather than as a pro-glacial lake of immense dimensions.

It seems probable that the erosion of the col-gully represents part of the same glacial sequence as the glaciogenic deposits occupying the Col; certainly no evidence has been found to the contrary.

The precise nature of the melt-water processes responsible for the erosion of the col-gully is open to debate and its elucidation depends upon the interpretation of the glacial sequence at the Col. It is clear from the stratigraphy that the main morphology of the col-gully was already largely in existence when the infilling ice-contact and glacio-lacustrine deposits collected within it. The col-gully might similarly represent erosion in a pro-glacial environment. An ice sheet advancing into the Yeo Lowlands might be anticipated to impound a series of small lakes as it impinged against Failand Ridge. Eventual overflow from such a small lake might be expected to erode a gully through the lowest point of the Col. Local advance and consequent elevation of the ice surfaces may have caused the edge of the ice sheet to enter the gully and cause the ice-contact deposits to accumulate within it. Subsequent continued advance may have resulted in the ice sheet over-riding the Col, with the till (unit 5) being deposited upon further advance, or later downwasting.

An alternative explanation can be envisaged (see Fig. 9) which would relate the col-gully and associated infill deposits to a downwasting, as opposed to an advancing ice mass. Clapperton (1968) has described how during the downwasting of an ice mass the increasing influence of the underlying relief might cause en- and sub-glacial drainage progressively to migrate towards, concentrate in and erode into areas where cols occurred in buried topographic ridges beneath the ice. This process would also explain satisfactorily the form and location of the Court Hill col-gully. Tills may have started to collect on the rock surfaces upon emergence of the relief above the ice mass, and the process envisaged by Mannerfelt (1945) might ensue. Surface water draining both from the emergent divides and ice masses may be added to pre-existing sub- and englacial drainage, perhaps greatly enlarging the col-gully. Ice-contact deposits then collected in the southern part of the col as the ice mass melted. A similar downwasting ice mass in the Vale of Gordano may have held up the small lake in which the gritty sand (unit 1) collected. The overlying till (unit 5) might owe its position to final movements of debris-rich material from the ice surface and adjacent hill-

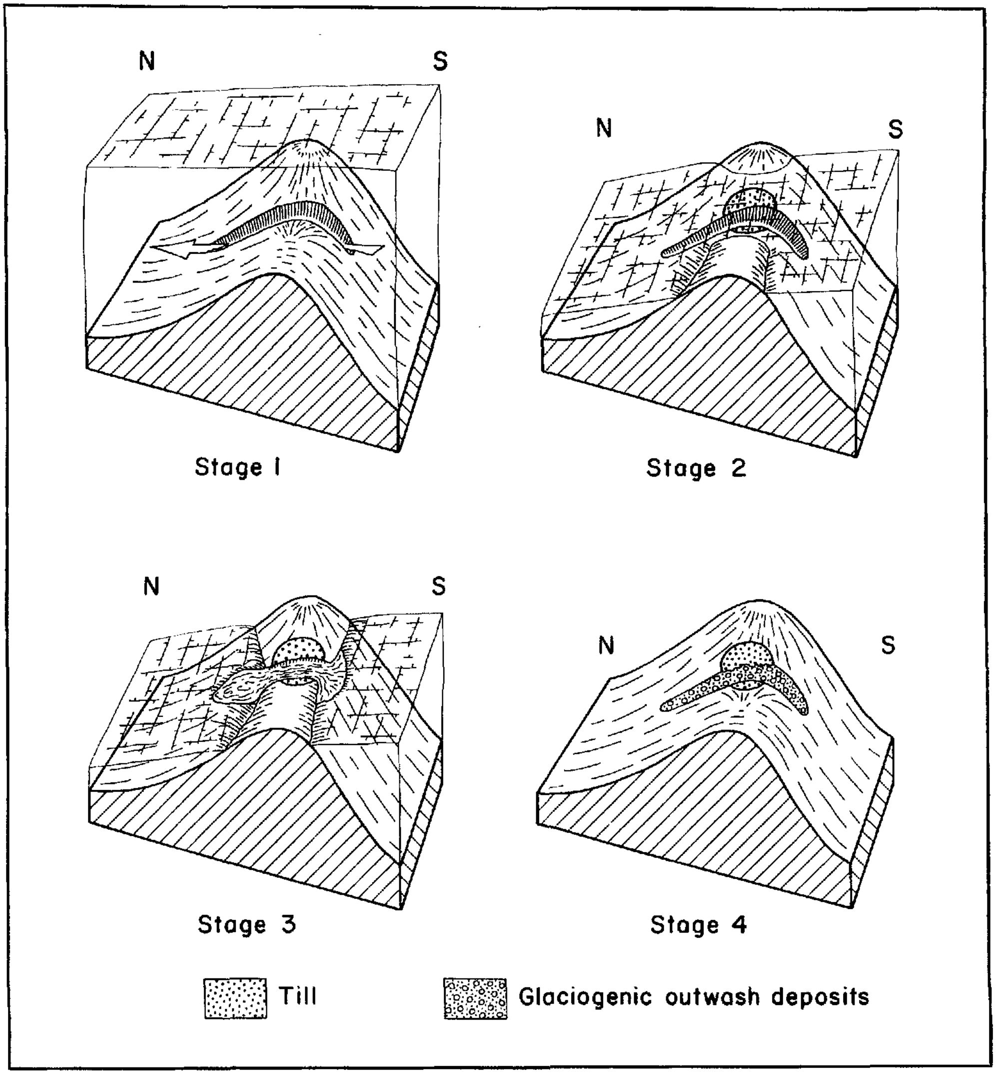

Fig. 9. Suggested stages in the development of the col-gully.

slopes. Extensive modification of unit 5 by quarrying, compaction, and grading renders as unreliable a verification by stone-orientation studies.

There is insufficient evidence available as yet to establish conclusively which of these two types of interpretation (or indeed the many possible combinations between them) actually occurred. However, the limited evidence available at present suggests the downwasting concept explains more features and presents fewer difficulties. For example, it would be surprising if an advancing ice sheet had not later gouged or flushed out previously deposited glaciogenic deposits in the col-gully, even if they were in a frozen state. The downwasting ice-mass concept would more readily provide an area of ponded water in the north of the col-gully at the same time as water and sediment were discharging into it from the higher ice mass to the south. An advancing ice sheet, however, entering the Vale of Gordano, might be anticipated to arrive after the ice from the Yeo Basin had completely crossed the col, hence too late to support a small ice-marginal lake in the vicinity of the Col. This is assuming that the hill

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ranges which almost encircle the Vale of Gordano (Fig. 1) were sufficiently high to have significantly delayed entry of an ice mass into the Vale.

Downwasting would also explain more satisfactorily the functional relationships between the cols and through-valleys of Failand Ridge at the time of glaciation. A logical downwasting sequence would suggest the progressive abandonment by sub-, en- and supraglacial melt-water drainage of the higher cols in favour of either the adjacent much lower and deeper East Clevedon Gap (Fig. 1) with its rock floor at least −6 m O.D., or through the buried channel system of the Yeo Lowlands to the south, described by Gilbertson and Hawkins (in press). If ice-sheet advance is invoked to explain the glacial sequence, it is difficult to understand why a pro-glacial lake, whose overflow was of sufficient dimensions to erode the Court Hill col-gully, should not have initially used, eroded, and later continued to overflow through the adjacent East Clevedon Gap. If the impounded ice-marginal lake was of such small dimensions that it could only overflow through the Col, it is unclear why such a large col-gully should have been eroded.

Although far from conclusive, the available evidence indicates the col-gully and glacial sequence are currently best interpreted in terms of a downwasting ice mass. The dimensions of the col-gully point to the importance of sub- and englacial drainage (perhaps aided by supraglacial drainage) in interpreting the gorge and valley landforms of the area.

Dating

Unfortunately, neither biogenic deposits nor weathering layers have been found which would have facilitated dating of the Court Hill site. Several lines of evidence suggest the glaciogenic features may be provisionally regarded as Wolstonian or older in age. First, they pre-date one phase of aeolian periglacial activity which is almost certainly of Devensian age. All the presently published evidence indicates that Devensian ice sheets did not cross the Bristol Channel/Severn Estuary into Somerset and Avon (Bowen, 1970, 1973; Gilbertson and Hawkins, in press; Gilbertson, unpublished). At nearby Kenn (Fig. 1), glaciogenic deposits of a very similar composition to those at Court Hill have been identified as probably Wolstonian in age, since they are overlain by interglacial deposits referred to the Ipswichian interglacial (Gilbertson and Hawkins, in press).

Limits of glaciation

The Quaternary deposits and landforms in Court Hill Col demonstrate that a considerable thickness of ice entered the coastal lowlands east of the Severn Estuary. The very localized survival of glacial deposits in this area makes reconstructing the limits of this glaciation difficult. Glacial drifts, provisionally attributed to the Wolstonian ice sheet, have been located in the Yeo Lowlands at Kenn and Yatton (Gilbertson and Hawkins, in press) and Churchill (Hawkins, 1972). The origins and ages of the drifts on the hill summits at about 100 m in the Bristol area (Davies and Fry, 1929) are more difficult to establish because of the sparsity of exposures. They are characterized by the presence of Greensand chert and Chalk flints, non-local rock types which are very common in the glacial deposits at Court Hill and Kenn. Drift deposits at Moat House Farm, Ashton Park, Leigh Wood, and Portishead Down reaching up to 122 m O.D. (Fig. 1) have been examined recently and interpreted as glacial in origin (Colbourne and others, 1974). This explanation is supported by the observation that the ice responsible for the glaciogenic deposits on the Col slopes at Court Hill must have attained a greater elevation than the 80 m O.D. reached by the "till" (unit 5). Consequently, it may be suggested that these high-level drifts may represent remanié tills or outwash deposits, and hence give a general indication of the area likely to have been occupied by ice sheets.

These observations support a glacial interpretation of many of the valleys and gullies of the Bristol district which, like the Court Hill col-gully, have anomalous relationships with both topography and geology. Typical are the col-gullies or valleys at Tickenham Hill, Cadbury Hill, Henbury, Providence and Rickford, which Hawkins and Kellaway (1971) and Hawkins (1972) suggested were only explicable as the results of erosion by glacial melt water. Their suggestion is supported by the nature and elevation of the find at Court Hill Col. As yet, no deposits or landforms are known in the area south-east of Bristol which would help establish the landward limit of glaciation. Since, however, there is clear evidence that the surface ice sheet reached at least 85 m O.D., and good reason to believe that on Failand Ridge it reached more than 120 m O.D. (Hawkins, 1972; Colbourne and others, 1974), there is every reason to believe that ice of this thickness would only have been stopped on its landward advance by hill masses such as the Mendips (250–300 m O.D.), Broadfield Down (180 m O.D.), Dundry (200 m O.D.) and the west-facing scarp of the Cotswolds.

Drainage diversions

The penetration of an ice sheet about 85–125 m thick into the Yeo Lowlands is likely to have been impeded by the constriction of the steep-sided vale in the Ubley–West Harptree district, giving credence to the views of Hawkins and Kellaway (1971) and Hawkins (1972) that the River Chew may have abandoned a seaward route through the lowlands and adopted a new easterly course as a result of the damming of the normal surface drainage (see Frey (1975) for an alternative view).

There is evidence for the former existence of small ice-marginal lakes in the area at Court Hill and Kenn, but as yet no evidence of a lake the size of Lake Maw. Indeed, the Severnside area may have been occupied by ice rather than water, introducing the erratic-rich drifts beneath the Severnside Lowlands, described by Morgan (1887).

Whilst this discussion has not conclusively established that the Avon Gorge at Clifton was formed by sub-, en- or supraglacially draining melt water, there now appears to be good circumstantial evidence indicating that it at least acted as a melt-water drainage channel.

Erratic-rich drifts and gorges on the high-level plateau surfaces of the Cotswolds above Bath, and on the Mendips, have been identified as the result of a much older and more widespread glaciation (Hawkins and Kellaway, 1971; Kellaway, 1971; Kellaway and others, 1971; Hawkins, 1972). This older glacial episode is at the moment distinguished in the Bristol area on the basis of the topographic location of its drifts. These lie on hill summits at and above 200 m, whereas the drifts tentatively associated with the Kenn and the Court Hill glaciation appear to lie in broad valleys or on plateaux eroded into and well below the higher drift-covered surfaces.

Conclusions

The Pleistocene deposits and landforms recently exposed in excavations at Court Hill Col demonstrate that an ice sheet penetrated into the lowlands east of the Severn Estuary, advancing inland from the Bristol Channel and Severn Estuary area.

The surface elevation of the ice sheet at Court Hill was at one stage at or greater than 85 m O.D. Other local evidence indicates an altitude of at least 125 m. The ice probably occupied most of the western lowlands of Avon County to Broadfield Down, Dundry Hill, and the Mendips, although there is at present no evidence for the eastern extent, which may have been at the Cotswolds scarp. On decay of the ice sheet, sub- and englacial melt waters' flow became concentrated in cols in the underlying relief where it eroded large col-gullies into the base of the pre-existing cols.

Several col-gullies, other than the best known at Court Hill, occur in Failand Ridge above Tickenham and Nailsea. Through-valleys such as the East Clevedon Gap, Rickford Valley, and Henbury Gap are attributable to sub- and englacial melt-water erosion. It is probable also that a significant part of the buried channel system of the coastal lowlands was similarly eroded by subglacial melt-water drainage. There is little evidence for the former existence of surface glacial melt-water spillways joining large glacially impounded lakes as proposed by Harmer (1907). Reversals of surface drainage by ice damming were caused in the Chew Valley. The Avon Gorge probably functioned, if indeed it did not originate, as a melt-water channel. On downwasting of the ice at Court Hill, stratified ice-marginal deposits accumulated on an ice slope leading into a small ice-marginal lake.

Apart from the nearby village of Kenn, other drift deposits in the area are poorly exposed. Future major excavations, however, may well reveal glacial deposits of the dimensions of those at Court Hill. The deposits and landforms at Court Hill, Kenn and below approximately 150 m are provisionally referred to as the Wolstonian or Anglian stage of the Pleistocene. The Devensian stage at Court Hill, and in much of the western part of Avon County, is characterized by the deposition of a blanket layer of periglacial cover sands. All these glacial and periglacial features are probably younger than another set of possible glacial deposits and landforms which occur at and above 200 m O.D. on parts of the Cotswold and Mendip Hills.

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INSTRUMENTS AND METHODS

THE DIFFICULTIES OF MEASURING THE WATER SATURATION AND POROSITY OF SNOW

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Abstract. Liquid saturation and porosity control most of the important material properties of wet snow, hence accurate measurements of these two parameters is of the utmost importance for both field research and glaciological applications. For example, the movement of liquid water through snow is highly sensitive to the volume of water present and accurate measurements of the water saturation are necessary in order to infer the temporal and spatial variations in the flow field. Nevertheless, most of the instruments in use are not capable of making accurate determinations of saturation.

An error analysis shows that only direct measurements of the liquid volume can provide accurate values of water saturation, hence the melting calorimeter is inherently inaccurate. While centrifuges extract some of the liquid for direct measurement, there is always some residual liquid left depending on the grain size and structural parameters of the ice matrix. Therefore, some uncertainty exists over the interpretation of the data obtained from centrifuges. High-frequency capacitance probes can be used either *in situ* or on the surface and are very sensitive to the volume of liquid present. Capacitance probes are by far the best of the available devices. Remote-sensing techniques, like the active microwave system, require more development for use in operational forecasting schemes and as research tools.

Résumé. *Les difficultés de la mesure de la teneur en eau et de la porosité de la neige.* La teneur en liquide et la porosité commandent la plupart des principales propriétés physiques de la neige mouillée; il en résulte que des mesures précises de ces deux paramètres sont de la plus haute importance pour la recherche sur le terrain comme pour les applications glaciologiques. Par exemple, le mouvement de l'eau liquide à travers la neige est extrêmement sensible au volume d'eau présent et des mesures précises de la teneur en eau sont nécessaires pour en déduire les variations dans le temps et dans l'espace du flux d'eau. Cependant, la plupart des appareils de mesure utilisés ne sont pas capables de donner des mesures précises de la teneur en eau.

Une discussion des erreurs montre que seules des mesures directes de volume liquide peuvent donner des valeurs précises de la teneur en eau, puisque le calorimètre de fusion est congénitalement imprécis. Lorsqu'on extrait l'eau liquide par centrifugation pour une mesure directe, il reste toujours une certaine quantité de liquide dans la neige variable selon la dimension des grains et les paramètres structurels de la matrice de glace. Par conséquent, il y a toujours une certaine incertitude sur l'interprétation des résultats obtenus par centrifugation. Les sondes à capacité à haute fréquence peuvent être utilisées soit *in situ* ou en surface et sont très sensibles au volume d'eau liquide présent. Les sondes à capacité sont, de loin, les meilleures parmi les appareils disponibles. Les techniques de télédétection, comme les systèmes à micro-ondes actives, demandent à être perfectionnées avant d'être utilisables en prévision opérationnelle et comme outils de recherche.

Zusammenfassung. *Die Schwierigkeiten bei der Messung der Wassersättigung und der Schneeporosität.* Flüssigkeitsgehalt und Porosität sind ausschlaggebend für die meisten wichtigen Materialeigenschaften nassen Schnees; genaue Messungen dieser beiden Parameter sind daher äusserst wichtig für die glaziologische Forschung wie für deren Anwendung. So ist zum Beispiel die Bewegung flüssigen Wassers durch Schnee stark abhängig vom vorhandenen Wasservolumen und genaue Messungen der Wassersättigung sind notwendig, um auf die zeitlichen und räumlichen Schwankungen des Fliessfeldes schliessen zu können. Trotzdem lassen sich mit den meisten gebräuchlichen Instrumenten keine genauen Bestimmungen des Sättigungsgrades vornehmen.

Eine Fehleranalyse zeigt, dass nur direkte Messungen des Flüssigkeitsvolumens genaue Werte der Wassersättigung liefern; das Schmelz-Kalorimeter ist deshalb von Hause aus ungenau. Wenn Zentrifugen auch einen Teil der Flüssigkeit zur direkten Messung verfügbar machen, so bleibt doch stets je nach Korngrösse und Struktur des Eisgefüges ein gewisser Rest zurück. Deshalb herrscht bei der Interpretation von Zentrifugendaten eine gewisse Unsicherheit. Hochfrequente Kapizitanzsonden können entweder *in situ* oder an der Oberfläche verwendet werden; sie besitzen hohe Empfindlichkeit für vorhandene Wassermengen. Kapazitanzsonden sind bei weitem die besten verfügbaren Geräte. Fernerkundungsverfahren, etwa mit aktiven Mikrowellensystemen, bedürfen noch der Fortentwicklung zur Nutzung für operationelle Vorhersagen und als Forschungsmittel.

SYMBOLS

c_0	mass of solution introduced
$E(x)$	relative error in parameter x
F	liquid-water content, liquid volume divided by total volume
g	acceleration due to gravity
g'	acceleration due to centrifuging
k	intrinsic permeability of snow
L	equivalent length of centrifuged sample
L'	length of centrifuged sample
m_0	molal concentration of sodium hydroxide
m_s	mass of snow sample
r	mean radius of centrifuge
S_m	mobile water, $S_w - S_{wi}$
S_w	water saturation, liquid volume divided by pore volume
S_{wi}	irreducible water saturation
S^*	effective water saturation, $(S_w - S_{wi})/(1 - S_{wi})$
t	equivalent time for a sample drained by gravity
t'	time of centrifuging
T	temperature
u	flux of water, flow through a unit area per unit time
v_a, v_i, v_s, v_w	volumes of air, ice, sample, and liquid
α	5.47×10^6 m^{-1} s^{-1}
β	molal temperature depression constant of the dissolved substance
ρ_i, ρ_s, ρ_w	densities of ice, snow, and water
ϕ	porosity, pore volume divided by total volume
ω	angular velocity of centrifuge

1. INTRODUCTION

There are many reasons for wanting to know the liquid-water saturation and porosity of snow. Every investigation of the snow cover uses information such as the snow density, state of metamorphism, liquid-water storage capacity, and/or liquid-water transmission rate. Furthermore, virtually all of the important material properties of snow are related to its density and history of liquid saturation. Two of the most important pieces of information about snow, either on the scale of kilometers or millimeters, are water saturation and porosity. These two pieces of information are necessary for even a crude approximation of such varied properties as snow strength and water flow rates. Accordingly, much attention has been given to developing devices for determining the liquid-water content and density of snow. While some of these devices work rather well (see Table I), others can be shown to be unacceptable conceptually. Because of the widespread use of devices which fall into the latter category, all of the popular devices are analyzed here in order to identify their limitations.

TABLE I. TYPICAL CALCULATION ERRORS WHEN $E(S_{wi}) = E(v_s) = E(m_s) = 0$

Method	*Measured values*	$E(\phi)$	$E(S_w)$	$E(u)$
i	v_a, v_s, m_s	$10.2E(v_a)$	$52.4E(v_a)$	$376E(v_a)$
ii	v_i, v_s, m_s	$E(v_i)$	$5.1E(v_i)$	$36.6E(v_i)$
iii	v_w, v_s, m_s	$0.16E(v_w)$	$0.84E(v_w)$	$6E(v_w)$
iv	v_w, v_i, v_s	$E(v_i)$	$E(v_w)+E(v_i)$	$13.6E(v_i)+5.6E(v_w)$

Snow hydrologists have been concerned principally with the "free-water content F" of snow, i.e. the fraction of the total volume occupied by the liquid phase. For most purposes it is more meaningful to separate the free-water content into its two component parameters, the liquid-water saturation S_w and porosity ϕ. These three are related by

$$F = S_w \phi. \tag{1}$$

Since it is necessary to determine both S_w and ϕ for virtually every wet-snow problem, F can be calculated from known values of S_w and ϕ if desired.

Neglecting the weight of the air phase,

$$\rho_s = \rho_i(1-\phi)+\rho_w \phi S_w, \tag{2}$$

relates the snow density ρ_s to the ice mass per unit sample volume plus the liquid mass per unit sample volume. Thus the snow density can also be calculated directly from known values of S_w and ϕ. The advantage of using S_w instead of F can best be illustrated by the flux–concentration relationship for water (Colbeck and Davidson, 1973),

$$u = \alpha k S^{\star 3}, \tag{3}$$

where

$$S^\star = (S_w - S_{wi})/(1-S_{wi}), \tag{4}$$

S_{wi} is the irreducible water content which cannot be removed by gravity drainage, α is a constant, the intrinsic permeability k is a function of only the porous matrix, and $S^\star$ is a function of only the mobile liquid fraction in the pore volume.

2. Calculation errors for S_w and ϕ

There are four fundamental equations; two describe the mass and volume balances and two are definitions (neglecting the air mass):

$$v_s = v_i + v_w + v_a, \tag{5}$$

$$m_s = \rho_w v_w + \rho_i v_i, \tag{6}$$

$$S_w \equiv v_w/(v_a+v_w), \tag{7}$$

$$\phi \equiv (v_a+v_w)/v_s. \tag{8}$$

These four equations contain seven variables hence three pieces of information must be supplied. Usually the sample mass and volume are determined independently, in which case either the air, water, or ice volume must be measured directly. If the sample mass or volume is not determined independently, then two other quantities would have to be measured. The errors inherent in four of the ten possible combinations of the volume and mass variables are analyzed here. First, the three possible cases which include v_s and m_s are considered. Then, the optimum combination for *in situ* sampling—v_s, v_i and v_w—is analyzed:

(i) *Measure v_a directly*

This procedure would be attractive for a destructive sampling technique because phase changes during sample preparation would not matter except that the solubility of air in water would have to be considered. Unfortunately, in order to calculate S_w and ϕ, it is necessary to divide by the difference between two large numbers, a procedure which inevitably leads to large errors. Using the definition of relative error,

$$E(x) \equiv \left|\frac{\mathrm{d}x}{x}\right|, \tag{9}$$

and assuming no error in the measurements of m_s and v_s,

$$E(\phi) = \frac{v_a \rho_w}{(\rho_w - \rho_i)(v_a + v_w)} E(v_a), \tag{10}$$

and

$$E(S_w) = \frac{v_a}{\rho_w - \rho_i}\left(\frac{\rho_i}{v_w} - \frac{\rho_w}{v_a + v_w}\right) E(v_a). \tag{11}$$

For a typical case where $v_i = 500 \times 10^{-6}\ \text{m}^3$, $v_w = 75 \times 10^{-6}\ \text{m}^3$, $m_s = 0.533$ kg, $v_s = 10^{-3}\ \text{m}^3$ and $v_a = 425 \times 10^{-6}\ \text{m}^3$,

$$E(\phi) = 10.2E(v_a), \tag{12}$$

and

$$E(S_w) = 52.4E(v_a). \tag{13}$$

These large values render this approach useless unless extremely precise measurements of v_a are possible. It is important to note that $E(S_w)$ increases rapidly as S_w approaches S_{wi}.

(ii) *Measure v_i directly*

While this method is widely used because of its simplicity, large errors are associated with calculating S_w from the measured quantities. Again assuming no error in the measurements of m_s and v_s,

$$E(\phi) = \frac{v_i}{v_s - v_i} E(v_i), \tag{14}$$

and

$$E(S_w) = \left(\frac{v_i}{v_s - v_i} - \frac{\rho_i v_i}{m_s - v_i \rho_i}\right) E(v_i). \tag{15}$$

Note that the liquid saturation only enters this equation through the denominator, $m_s - v_i \rho_i$. As the liquid saturation decreases, $m_s - v_i \rho_i$ decreases and a large error occurs in the calculation.

For the sample case,

$$E(\phi) = E(v_i), \tag{16}$$

and

$$E(S_w) = 5.1E(v_i), \tag{17}$$

but $E(S_w)$ is much higher at lower values of saturation. As shown on Figure 1, $E(S_w)/E(v_i)$ approaches negative infinity as S_w approaches zero.

While this method may be useful for calculating the snow porosity, it will inevitably lead to large errors in the calculated value of water saturation unless very precise measurements of v_i are possible. As shown later, if this method were used to calculate or infer variations in the flow field of liquid water, a great deal of uncertainty would be associated with the results.

(iii) *Measure v_w directly*

This is the principle of many common devices and, since the error analysis shows that this method is most likely to produce consistent and accurate results, it is analyzed in more detail. We find

$$E(\phi) \leqslant \frac{v_i}{v_s \phi} E(v_s) + \frac{m_s}{\rho_i v_s \phi} E(m_s) + \frac{\rho_w v_w}{\rho_i v_s \phi} E(v_w), \tag{18}$$

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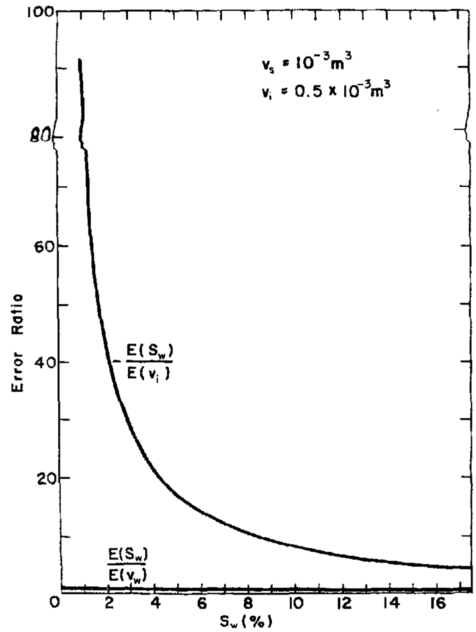

Fig. 1. The ratios of error in calculating S_w to the error in measuring v_i and v_w are shown for methods (ii) and (iii) respectively (assuming $E(m_s) = E(v_s) = 0$, $v_s = 10^{-3}$ m³ and $v_i = 0.5 \times 10^{-3}$ m³). For method (ii) the ratio is very large for all values of S_w and approaches negative infinity as S_w vanishes. For method (iii), however, the ratio reaches an upper limit of unity as S_w vanishes and is rather insensitive to the value of S_w over the common range of saturations.

and

$$E(S_w) \leqslant \left(\phi^{-1} - \frac{m_s}{\rho_i v_s \phi}\right) E(v_w) + \phi^{-1} E(v_s) + \left(\frac{m_s}{\rho_i v_s \phi}\right) E(m_s), \tag{19}$$

where the inequality sign is used to show that we only know the upper limit of the error. Unlike the two previous methods, as the volume of water approaches zero, the error in S_w approaches a small upper limit instead of becoming infinite (see Fig. 1). For the sample case,

$$E(\phi) \leqslant 0.164 E(v_w) + 1.0 E(v_s) + 1.16 E(m_s), \tag{20}$$

and

$$E(S_w) \leqslant 0.84 E(v_w) + 2.0 E(v_s) + 1.16 E(m_s). \tag{21}$$

From these considerations it is clear that the direct measurement of the liquid is inherently a more accurate method of determining both S_w and ϕ (see Table I). The use of methods (i) or (ii) could only be justified if it could be shown that the error in measuring the volume of water directly was much greater than the error in measuring the volumes of ice or air. However, as shown later, the dielectric constant is very sensitive to the liquid volume, hence an accurate determination of S_w is possible.

For method (iii) the errors in calculating ϕ or S_w are more sensitive to the errors of measurement of m_s and v_s than of v_w. As shown by Equations (18) and (19), $E(\phi)$ and $E(S_w)$ increase inversely with the quantity $\rho_i v_s \phi$. For a given situation, the errors in S_w and ϕ decrease as sample size increases, thus demonstrating the importance of avoiding small samples.

(iv) *Measure v_w, v_i and v_s directly*

The use of a snow sampling kit to measure m_s and v_s is inherently undesirable because it is frequently necessary to use two samples to get one calculated value of S_w or ϕ. Direct measurements of v_w, v_i and v_s are most desirable. In principle these measurements are possible, for example, by an electromagnetic instrument which senses the volumes of solid and liquid in a volume of sample which is predetermined by the nature of the instrument. The errors in calculating S_w and ϕ are given by

$$E(\phi) \leqslant \frac{v_i}{v_s - v_i} \left(E(v_i) + E(v_s)\right), \tag{22}$$

and

$$E(S_w) \leqslant E(v_w) + \frac{v_s}{v_s - v_i} E(v_s) + \frac{v_i}{v_s - v_i} E(v_i). \tag{23}$$

For the sample case these errors are

$$E(\phi) \leqslant E(v_i) + E(v_s), \tag{24}$$

and

$$E(S_w) \leqslant E(v_w) + 2E(v_s) + E(v_i). \tag{25}$$

If the predetermined volume being sensed v_s is known accurately, this method should work well for a remote-sensing system which can accurately distinguish between the liquid and solid phases.

3. Calculation errors for u

The liquid-water content is most often measured to infer or calculate variations in the flow field of water. Because of the sensitivity of water flux to water saturation as shown by

$$u = \alpha k S^{\star 3}, \tag{3}$$

any error in the determination of $S^\star$ will be magnified in the calculation of u. The related error in flux is bounded by

$$E(u) \leqslant 3E(S^\star) + E(k). \tag{26}$$

Accordingly, errors inherent in the device used to determine S_w must be avoided as much as possible. The error in calculating flux could be significantly reduced by taking a large number of samples in order to determine S_w more accurately. This procedure is difficult to follow in practice, however, since destructive sampling procedures are inherently undesirable and *in situ* instrumentation is rarely used in multiple arrays at each level in a snow-pack. The question of lateral variability must also be addressed if multiple samples are to be taken from a single snow layer.

At worst $E(u)$ increases directly as $E(k)$ but, since direct measurements of intrinsic permeability are rarely made, k must be calculated from determinations of ϕ and the average grain size (e.g. Shimizu, 1970). From data given by Kuroiwa (1968), it can be shown that k increases as exp (15.9ϕ), which is a typical permeability–porosity relationship for a porous medium. Neglecting any error in the determination of the average grain size, the relative error in permeability is highly sensitive to $E(\phi)$ as shown by

$$E(k) = 15.9\phi E(\phi). \tag{27}$$

Again, the need to determine ϕ accurately is apparent.

From Equation (4),

$$E(S^\star) \leqslant E(S_m) + \frac{S_{wi}}{1 - S_{wi}} E(S_{wi}), \tag{28}$$

where the "mobile-water saturation" is given by

$$S_m = S_w - S_{wi}. \tag{29}$$

As a consequence, $E(S^\star)$ is most sensitive to the measurement of S_m since $S_{wi}/(1-S_{wi})$ is a small number. When the error in $S^\star$ is expressed in this way, the advantages of using a system which measures only the mobile water are apparent. This result suggests that for the purpose of calculating the flux of water, a controlled capillary withdrawal of the mobile liquid might be better than measurement of the total liquid saturation. For the determination of the material properties, however, the total liquid saturation is required.

$E(S^\star)$ can also be expressed as

$$E(S^\star) \leqslant \frac{S_w}{S_w - S_{wi}} E(S_w) + \frac{(1-S_w)\, S_{wi}}{(1-S_{wi})(S_w - S_{wi})} E(S_{wi}), \tag{30}$$

which shows that $E(S^\star)$ becomes very large as S_w approaches S_{wi}. Apparently, the error in calculating flux from measurements of S_w becomes very large as flux becomes vanishingly small.

From direct determinations of S_w, S_{wi} and ϕ, the error in calculating the flux of water is bounded by

$$E(u) \leqslant 15.9\phi E(\phi) + \frac{3S_w}{S_w - S_{wi}} E(S_w) + \frac{3S_{wi}(1-S_w)}{(1-S_{wi})(S_w - S_{wi})} E(S_{wi}). \tag{31}$$

The upper bound of $E(u)$ is approximated at large fluxes by

$$E(u) \leqslant 5E(\phi) + 6E(S_w) + 2.5E(S_{wi}), \tag{32}$$

and the error increases with decreasing flow rates. Again, the need to determine the liquid content accurately is apparent. Likewise, the need to make an accurate porosity determination is shown.

If method (ii) is used to determine the liquid saturation, even in the absence of errors in the measurements of m_s, v_s or S_{wi}, for large flow rates

$$E(u) = 36.6E(v_i), \tag{33}$$

and $E(u)$ increases rapidly as u decreases.

This large error in the calculated value of flux precludes the possibility of getting a meaningful estimate of the flow field from measurements of the ice volume and explains why reproducible results have not been generally obtained with a melting calorimeter.

If method (iii) is used and the volume of liquid is sensed directly, at large flow rates

$$E(u) \leqslant 6.03E(v_w) + 19.2E(v_s) + 15.7E(m_s). \tag{34}$$

Even in the absence of error in measuring v_s, m_s or S_{wi}, the error in calculating the flux of water from the measured volume of water is six times greater than the error in the measurement itself. Assuming equal errors in the measurements of v_w and v_i, the advantage of working directly with the volume of water is apparent. It is important, however, that the volume of ice is generally about six times greater than the volume of liquid hence its error of measurement might be somewhat lower. These errors must be considered, for example, in choosing between the melting and freezing calorimeters.

If method (iv) is used with a remote-sensing system, the error in calculating flux would be

$$E(u) \leqslant 13.6E(v_i) + 5.63E(v_w) + 19.2E(v_s). \tag{35}$$

Clearly this method compares favorably with any of the others although large errors could still occur in the calculation unless very accurate measurements are possible.

4. Destructive sampling techniques

With this background, a systematic discussion of the various types of "saturometers" can be made. Excluding those which cannot give accurate results *a priori*, we consider two types—those which must take a sample from the snow cover and those which can get the necessary information *in situ*.

4a. *Freezing calorimeter*

Radok and others (1961) discussed the inherent advantage of the freezing calorimeter. They point out that in calculating "snow quality" from measurements with a melting calorimeter, it is necessary to take the small difference between large numbers, hence large errors are possible. In the freezing calorimeter, however, the calculation is different and the errors are not likely to be so large. This conclusion is identical to that derived above for techniques which measure v_w versus those which measure v_i.

Leaf (1966) reports achieving an accuracy of 1% liquid by weight or $\frac{1}{2}$% by volume with this technique. This suggests that S_w could be calculated to within an accuracy of 1%, a feat which would only be possible if there were essentially no errors in measuring v_s and m_s and if v_w was calculated to within 1% from the calorimetry data. The latter seems unlikely in view of the difficulty of handling the snow samples without causing any phase changes, but nevertheless this method is promising and hopefully further refinements will be made. The inherent advantage of a direct measurement of v_w probably compensates for the fact that freezing is more difficult than melting. Perhaps this method's major disadvantage is the fact that it is a destructive sampling technique, hence some disturbance of the flow field will necessarily be associated with its use.

4b. *Centrifuges*

Much attention has been given to the use of centrifuges in soil physics, petroleum reservoir engineering, and snow hydrology. The attractive feature of a centrifuge is that large accelerative forces are exerted on the fluids, thereby decreasing the time necessary to drain a porous sample. Stallman (1964) shows that

$$L' = Lg/g', \tag{36}$$

$$t' = t(g/g')^2, \tag{37}$$

and

$$g' = \omega^2 r, \tag{38}$$

where primed length L', time t' and acceleration g' are for the centrifuge's frame of reference and unprimed L and t are their equivalents in a sample subjected only to Earth's gravity g. When $g' = 1\,000g$, a 10 mm long sample centrifuged for 40 min is equivalent to the drainage of a 10 m column for 76 years *provided the bottom of the sample remains in contact with the liquid during the 40 min of centrifuging*. Unfortunately these scaling laws do not apply to the centrifuges currently in use by snow hydrologists because the samples do not remain in contact with the extracted liquid so there is some uncertainty about how the snow centrifuge should be scaled. Furthermore, Slobod and others (1951) state that displacement occurs only down to the "connate value" or irreducible water saturation. If Slobod and others are correct, only the mobile water can be extracted by this method and the total liquid-water content cannot be found by centrifuging! This does not necessarily rule out the use of centrifuges since, even if only the mobile component of the water were removed and measured, this could be useful information. It must be noted, however, that centrifuging snow samples at 60 revolutions per second can cause some consolidation. The time of consolidation effectively increases by a factor of $(g'/g)^2$ (Terzaghi, 1942).

Terzaghi (1942) described centrifuges by noting that, if the accelerative force is increased N times, the body forces are increased to NS_{W} per unit volume while the retentive forces remain constant. Therefore the height of capillary rise is decreased to $N^{-1}h_{\mathrm{c}}$, hence the capillary end effect is reduced. Terzaghi also states that this scaling is invalid when the water is discontinuous because the weight of the particles becomes small enough that the surface tension can balance it. Contrary to Slobod and others, Terzaghi believes that the discontinuous moisture in centrifuging is less than the discontinuous moisture in gravity drainage, hence more than just the mobile water is extracted. Unfortunately, there are different interpretations of just how much liquid can be extracted by centrifuging. This question is more important for snow than for most other porous materials because of the limitations placed on the centrifuging of snow by the problems of melting and compacting during the process. To provide some insight into the question of how much liquid can be extracted by centrifuging, samples of hydrophilic glass beads were soaked with water and centrifuged under conditions similar to those possible in a field situation. Although glass beads were used in place of snow in these tests, the results are applicable to snow because all of the important parameters which affect the retention of water in snow were closely simulated in these tests. The samples were well soaked to ensure complete wetting and then hand-centrifuged for 60 s at about 7 revolutions per second to reduce their liquid content to about that of a freely draining snow cover. The samples were then centrifuged by a machine for 60 s at 33.3 revolutions per second plus the 30 s necessary to accelerate and decelerate the machine. The residual water present following centrifuging, as determined by oven-drying the samples, was highly dependent on the size of the glass beads (see Fig. 2) over the range of grain sizes common in natural snow covers

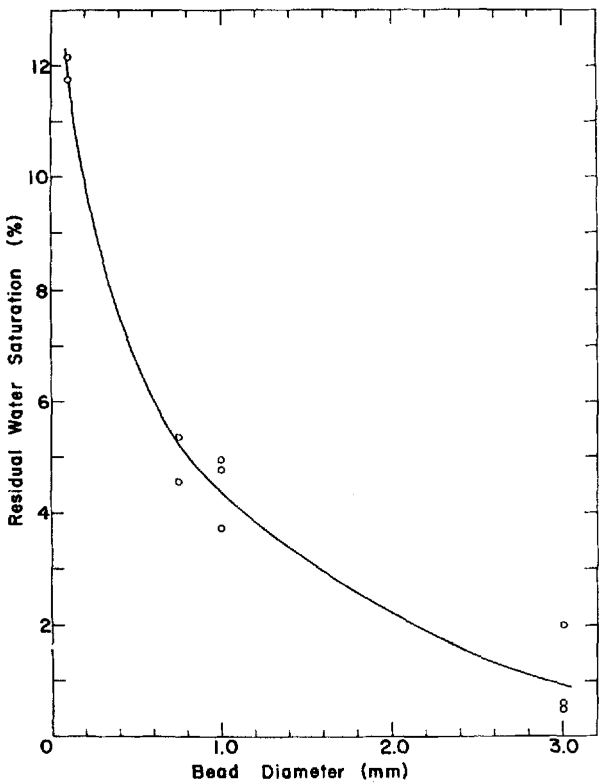

Fig. 2. The residual water saturation left behind after centrifuging decreases with increasing particle size. Much uncertainty is connected with the use of centrifuges because of this residual water.

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(Wakahama, 1968). This result shows that different amounts of water will be extracted from different parts of a snow pack due to the occurrence of different grain sizes. Therefore, even as a *relative* measure of the amount of water present and the amount of water flowing, the centrifuge can give misleading results.

In the past Yosida (1967) and LaChapelle (1956) have noted the partial retention of water by snow samples during centrifuging. Hopefully, the results given here will discourage the widespread use of centrifuges in snow hydrology. Centrifuges have often been used as a measure of the spatial variability of the flow field of water in snow (e.g. Langham, 1974) but as shown by Equation (12), the error in flow is six times as large as the error in S_w. Unfortunately, the error in S_w as determined by centrifuging is affected by such things as grain size, melting, compaction, and time and rate of spinning. Under any circumstances, some water will be retained by the snow sample, a fact which precludes the quantitative use of the information obtained from a centrifuge. LaChapelle (1956) showed that some qualitative information can be obtained with a centrifuge, but the use of centrifuges for purposes such as determining the spatial and temporal variability of the flow field in normal snow covers is risky. Much of the infered non-uniformity in the flow field is simply due to the inherent limitations of the centrifuge.

4c. *Solution method*

Bader (1948) proposed a simple method in which a dilute solution of sodium hydroxide is added to a known quantity of wet snow and the temperature depression is measured. As long as the temperature depression is small (≈ 1.5 deg) and no significant errors occur in measuring the solution weight or sample weight, an accurate determination of the free water content F can be made. The relative error is

$$E(F) = \frac{\beta m_0 c_0 + 0.0062(m_s + c_0)\ T^2}{\beta m_0 c_0 - 0.0062(m_s + c_0)\ T^2 - c_0 T} E(T), \tag{39}$$

or, for a typical case described by Bader (1950),

$$E(F) \approx 1.5E(T). \tag{40}$$

Thus Bader's method offers a quick and easy alternative to calorimetry. Assuming accurate weight, volume and temperature measurements, useful information on the free water content should be obtainable. Probably the biggest error in this method would be introduced by inaccurate measurements of the molal concentration m_0 of the solution of sodium hydroxide. Neglecting other errors,

$$E(F) = \frac{\beta c_0 m_0}{TF} E(m_0), \tag{41}$$

or, for a typical case,

$$E(F) \approx 3.1E(m_0). \tag{42}$$

This shows that the solution must be prepared carefully.

5. Non-destructive measuring devices

While some of these destructive sampling techniques give reasonably accurate information about the liquid water in a sample, they all suffer the serious disadvantage of disturbing the flow field by the act of removing the sample. This fact precludes repeated sampling to determine the temporal variations at a point in a snow cover. Repeated sampling to determine the local spatial variations is also suspect because of disturbances to the flow field caused by the creation of new surfaces within the flow field. Accordingly, the use of *in situ* or remote-sensing devices is necessary to obtain the most useful information, and several such devices are reviewed here. Other possible methods, which have not yet been applied to snow, include nuclear magnetic resonance (NMR), time-domain reflectometry, Raman scattering (see Miller, 1972), and acoustic methods.

5a. *Dielectric devices*

The large contrast between the dielectric constants of liquid water and ice at megahertz frequencies has provided the basis for measuring the liquid-water content of various materials at least since the early 1930's when meters were used to determine the liquid content of wheat. Perhaps Gerdel (1954) was the first to apply these devices to snow using a capacitance probe operating at a frequency of 1.5 MHz. Although the dielectric constant of snow is very sensitive to small changes in the volume of liquid water present, it is difficult to make a strict interpretation of the dielectric constant of the solid–liquid–gaseous mixture because of the importance of shape factors on the contribution of each phase to the dielectric constant of the mixture.

Ambach and Denoth (1975) have recently improved the capacitance probes used in snow by designing an instrument which operates at frequencies up to 20 MHz. At higher frequencies, the effect of grain size is minimized but the snow density must be known to calculate the free water content from the dielectric constant (Ambach and Denoth, 1975). The need to measure the snow density, a distinct disadvantage of this approach, arises because the mixing formulae (used to account for the shape factors) require the use of the value of the dielectric constant extrapolated to an infinite frequency. The dielectric constant then has only a real part, hence the density must be determined separately in order to supply enough information to calculate both the porosity and liquid saturation. For the purposes of designing an *in situ* instrument, the lack of information about the imaginary part of the dielectric constant is very unfortunate. Nevertheless, the dielectric devices are useful instruments with a standard error of about 0.5% by volume (Thomas, 1966; personal communication from A. Denoth).

At microwave frequencies, the real and imaginary parts of the dielectric constant can be determined simultaneously, thus providing all of the information necessary to calculate the porosity and water saturation in a known volume of wet snow (Sweeny and Colbeck, 1974). Unfortunately, the sophisticated microwave equipment is not easily adapted to field situations, so dielectric measurement by capacitance (Ambach and Denoth, 1975) is probably the best available at this time. Perhaps the best hope for developing a system to obtain information about porosity and water saturation in wet snow is the active microwave system. Linlor and others (1974) describe the application of such a system for obtaining profiles of snow wetness. The ultimate system would determine profiles of solid and liquid contents in a snow cover, thus providing information about the degree of layering, propagation of melt-water waves, state of ripening, depth, etc.

5b. *Other devices*

The soil-water tensiometer of Richards and Gardner (1936) measures the negative gage pressure in the liquid phase in an unsaturated soil. In snow, the negative pressure, or "tension", is determined primarily by the liquid-water saturation and grain size. For a thoroughly wetted snow where the grain size is stable but the liquid saturation changes with the flux of water, tension measurements provide a direct indication of the flow rate (Colbeck, 1976). This correlation between flow rate and tension at a point provides a useful method for observing the flow field without providing any direct knowledge of the porosity or water saturation. For example, Wankiewicz (unpublished) used tensiometers to make extensive observations of the flow of water in a deep mountain snow-pack. Unfortunately, tensiometers are difficult to use in a large-grained, porous medium like snow, hence *in situ* measurements with a capacitance probe may be more successful. Also, capacitance probes are free from the freezing problems of tensiometers; the capacitor can be positioned before the seasonal snow falls and left in position during freeze–thaw cycles.

Methods currently in use to determine the water equivalent of snow include terrestrial gamma-ray surveys (Peck and Bissell, 1973) and nuclear profiling gages. The use of nuclear sources which are attenuated by the snow was introduced by Gerdel and others (1950) and refined by Smith (e.g. Smith and others, 1965). The resolution of these gages is approaching the point of development necessary to detect changes in the flow rate of water. However, even in the absence of any error in ϕ or S_{wi}, Equation (12) shows that $E(u)$ is typically six times as large as $E(S_w)$. If at most a ten per cent error in flow rate is desired, the liquid-water saturation would have to be determined to within a relative error of less than two per cent. Typically this means a measurement of total water mass accurate to within two parts per thousand, a feat which is beyond the capability of the nuclear gages currently in use.

6. Conclusions

Liquid-water saturation and porosity are two of the most basic pieces of information about a snow cover. These two parameters largely control such important properties as reflectivity, rheology, and water flow rates. Nevertheless, adequate methods for measuring these properties do not exist and further developments are necessary.

Besides the inherent limitations of destructive sampling techniques, the two most commonly used sampling devices have physical limitations. The centrifuge does not extract all of the liquid, a fact which has been frequently cited in the past. The problem with the centrifuge is that the water left in the sample is a complicated function of structural parameters such as grain size. For example, the water left behind in a sample of glass beads depends on the grain size over the range of sizes commonly observed for snow (see Fig. 2). This fact precludes the use of a centrifuge to infer accurately the flow field in a snow cover. From data taken by the melting calorimeter, both the water saturation and porosity can be calculated. While the calculated value of porosity may be fairly accurate, the calculated value of water saturation is generally highly inaccurate since any error in the measurements is magnified by the nature of the calculation for the liquid-water saturation (see Fig. 1). Accordingly, the use of either the melting calorimeter or centrifuge to determine the spatial or temporal variations in the flow field would be risky. What appear to be variations in flow might just be due to errors inherent in the methods.

Although the response of a tensiometer has been related experimentally to the liquid flow rate in snow, the high-frequency capacitors seem to offer more advantages for making *in situ* determinations of the liquid-water saturation for research studies. The advantages of the capacitance probes include easy coupling with the snow and a lack of freeze–thaw problems. Eventually efficient methods must be developed for interrogating the snow cover remotely in order to provide the necessary information for hydrological forecasting practices.

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TESTING OF RADIOMETRIC DETECTION OF AVALANCHE VICTIMS

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Abstract. A radiometer prototype has been constructed using the S band. In this band, a theoretical evaluation of the differences of temperature which are measurable by the radiometric method proves the influence of the parameters liquid-water content of snow and snow thickness.

When there is no victim, the apparent temperature of a snow layer depends on these parameters and presents important variations because of its natural inhomogeneity. In order to detect the local variation of the apparent temperature due to the presence of a victim, we must refer to an average local temperature obtained by a servo-system. Thus, we obtain a certain number of false alarms which can be suppressed only with a manual sounding. The search itself becomes very long and also requires a systematic exploration of the snow surface by parallel traverses spaced every 2 m. This exploration seems impossible on uneven ground such as a real avalanche area.

Résumé. *Essais de détection radiométrique des victimes d'avalanche.* Un prototype de radiomètre a été réalisé en bande S. Une évaluation théorique dans cette bande des écarts de températures mesurables par la méthode radiométrique met en évidence l'influence des paramètres teneur en eau de la neige et épaisseur de neige.

En l'absence de victime, la température apparente d'un manteau neigeux dépend des paramètres précédents et présente de grandes variations par suite de son inhomogénéité naturelle. Pour détecter la variation locale de température apparente due à la présence d'une victime, il faut se référer à une température locale moyenne, obtenue par asservissement. Il existe alors un certain nombre de fausses alarmes que l'on ne peut éliminer que par sondage manuel. La procédure de recherche devient très longue et nécessite en outre une exploration systématique du terrain par des allers et retours parallèles espacés de l'ordre de 2 m. Cette exploration semble impossible sur un terrain tourmenté comme l'est un terrain d'avalanche réel.

Zusammenfassung. *Die Suche nach Lawinenopfern durch Radiometrie.* Der Prototyp eines Radiometers im S-Band wurde gebaut. Eine theoretische Studie der Temperaturunterschiede, die in diesem Bande durch Radiometrie gemessen werden können, zeigt den Einfluss des Wassergehaltes und der Dicke der Schneedecke.

In Abwesenheit eines Lawinenopfers hängt die scheinbare Temperatur der Schneedecke von den obengenannten Parametern ab und weist infolge der natürlichen Unregelmässigkeit starke Schwankungen auf. Um die örtliche Veränderung der scheinbaren Temperatur zu messen, die durch ein verschüttetes Lawinenopfer bedingt ist, muss mas sich auf die mittlere lokale Temperatur beziehen. Dies geschieht mittels eines Regelsystems. Dabei ist eine gewisse Anzahl von falschen Alarmen nicht zu vermeiden und nur die Handsonde kann in solchen Fällen entgültige Klarheit verschaffen. Die Suche dehnt sich dadurch sehr lange aus, und das Lawinengebiet muss meanderförmig abgeschritten werden, wobei die Hin- und Hergänge im Abstand von etwa 2 Metern erfolgen. Diese Suchmethode scheint sich über unebenem Grund wie in wirklichen Lawinengebieten kaum verwirklichen zu lassen.

Introduction

In order to detect avalanche victims, we want to show by radiometry that there is thermal radiation of a human body buried in a snow layer, that is to say a dielectric area more or less absorptive, placed on the ground. The thermal radiation transmitted above the snow is collected by an antenna which changes this radiation into a slight noise characterized by an apparent antenna temperature.

In order to use antennas of not too large a size, it is necessary to work with high frequencies. On the other hand, the rather low radio frequencies are the only ones that can penetrate a compact, wet snow without excessive attenuation. Therefore a compromise is required. It seems that an interval of frequencies from 2 GHz to 8 GHz can be used.

Now, the detection of a slight noise signal can only be obtained with an ultra-sensitive receiver which uses a pre-amplifier with slight noise and high gain. Besides, the power of the useful noise that characterizes the presence of a thermal source becomes greater and greater as the band width of the receiver becomes wider.

The technology of today allows the production in the S band of a radiometer sufficiently reliable to detect a variation of apparent antenna temperature of about one kelvin (Liva, 1975).

In order to detect avalanche victims, the radiometer is composed of two similar, mechanically coupled antennas that can be moved about over the avalanche area which is to be explored.

A continuous electric signal is obtained at the radiometer output proportional to the difference between the apparent temperatures of the antennas. A servo-motor can cancel out the continuous signal voltage above so that an average local temperature of the snow layer can be defined.

The time constant chosen for the servo-system must be large enough to obtain for a speed of movement of about 0.3 m/s a signal that can be detected when there is a victim under a snow layer. A certain number of false alarms will also be obtained.

We present a model to calculate the difference of the apparent temperatures of antennas, pointing out the importance of the two parameters: water content of snow and thickness of snow (Liva, unpublished). This model can explain why there are false alarms, and can predict the signal detected when there is a human body under a snow layer.

Finally, a systematic study of detection on snowy ground can assess the feasibility of the application of the radiometric method for the detection of avalanche victims.

Model for calculating the difference of the apparent temperatures of antennas

The location of the antennas and of the human body buried in the snow is shown in Figure 1.

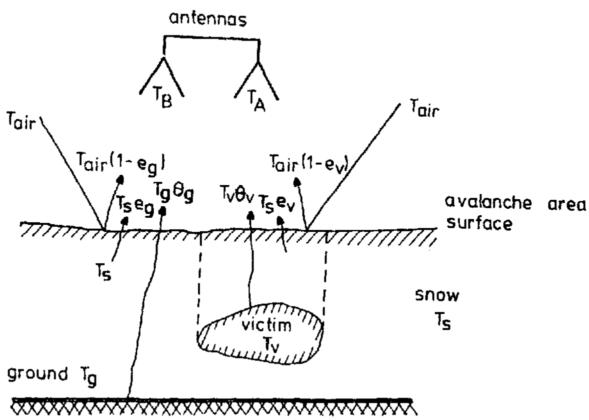


Fig. 1. Location of the antennas and of the human body in the snow. T_A, T_B are the temperatures of the antennas, T_{air}, T_s, T_v, the real temperatures of the three materials involved.

This supposes that the antennas are similar and directive enough so that only the A antenna is able to receive the thermal radiation from the victim. We also suppose that the irregularities of the avalanche area surface are slight compared with the working wavelengths.

We respectively call the temperatures of the air, snow, victim and ground, T_{air}, T_s, T_v, and T_g respectively.

The variation of the apparent temperatures of the antennas can be written:

$$\Delta T_{app} = T_A - T_B = (T_v - T_s)\,\theta_v + (T_{air} - T_s)(e_g - e_v) - (T_g - T_s)\,\theta_g,$$

where θ_v and θ_g represent the rates of transmission through the snow and e_v and e_g the emissivities of the snow above the human body or the ground.

In practice $T_g - T_s$ remains small so that the last term in the equation for ΔT_{app} can be neglected. The values of e_v, e_g and θ_v represented respectively on Figures 2, 3 and 4 are calculated from a plane model with three elements (Liva, unpublished).

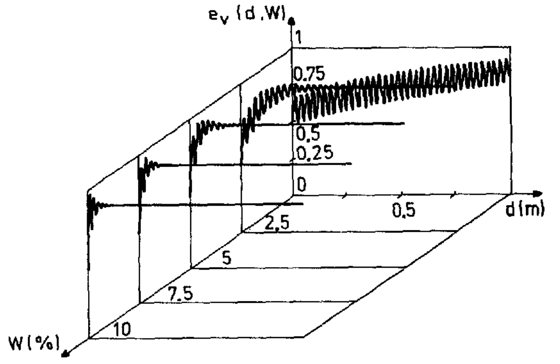


Fig. 2. Emissivity $e_V(d, w)$ at the air–snow interface of a homogeneous snow layer with a thickness d and a water content w in volume, overlying an infinite plane of muscular fibre at 4 GHz.

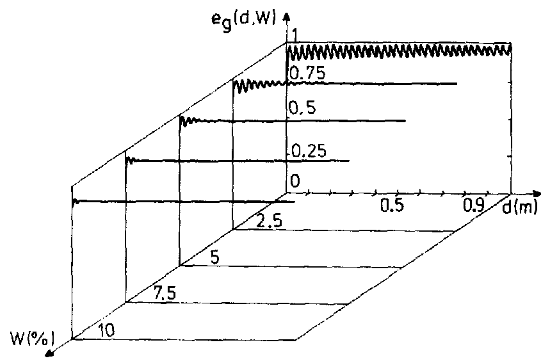


Fig. 3. Emissivity $e_g(d, w)$ of snow with thickness d and water content w in volume at 4 GHz. Snow layer on sandy, wet ground.

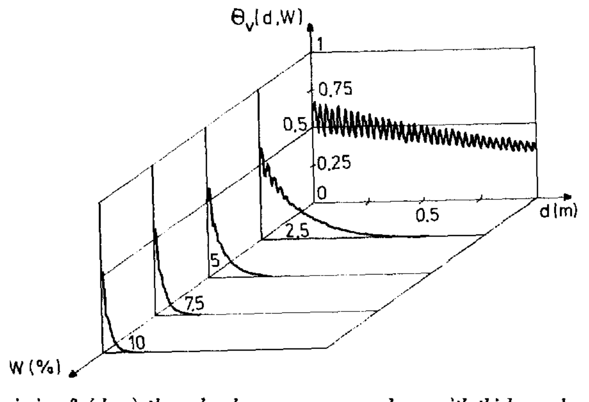


Fig. 4. Rate of power transmission $\theta_V(d, w)$ through a homogeneous snow layer with thickness d and water content w in volume, placed on an infinite plane of muscular fibre at 4 GHz.

The values given for ΔT_{app} in Table I are approximate to within a few degrees, and can also be cancelled out.

The difference between the apparent temperatures of the antennas is largest when one of the antennas is over the victim. When the antennas system is in action above the victim, ΔT_{app} undergoes an inversion passing through zero. The distance between the antennas is chosen to be about one metre.

The distance within which the system of antennas goes from one extremum of ΔT_{app} to the other depends on the size of the victim and is between 0.3 m and 1.5 m. A variation of ΔT_{app} of the same type can also be obtained when there is no victim, as a result of inhomogeneity of a snow layer. In that case, we can write:

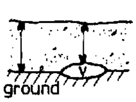
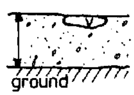
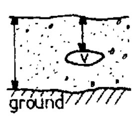
$$\Delta T_{\text{app}} = (T_{\text{air}} - T_{\text{s}})(e_{\text{gA}} - e_{\text{gB}}),$$

where e_{gA} and e_{gB} represent the emissivities of the snow beneath the A and B antennas respectively. If the emissivity of the snow presents local fluctuations on a scale of about one metre, we obtain a false alarm.

A logical device placed in the chain of treatment of the noise signal connected with ΔT_{app}, allows us to take into account only variations that could come from a victim.

TABLE I. ΔT_{app} VALUES, IN PRESENCE OF VICTIM

Elements: wet ground, muscular fibre, snow; $F = 4$ GHz; Conditions: $T_{\text{g}} = T_{\text{s}} = 273$ K, $T_{\text{v}} = 310$ K

Models d_g =		d_v =	w %	θ_{v}	e_{g}	e_{v}	ΔT_{app} $T_{\text{air}} =$ 73 K K	ΔT_{app} $T_{\text{air}} =$ 123 K K	ΔT_{app} $T_{\text{air}} =$ 223 K K
0.5 m		0.5 m	0	0.40	0.92	0.68	−33.00	−23.00	−9.00
			1	0.20	0.93	0.85	−8.60	−4.60	−0.60
			2	0.05	0.96	0.97	+3.85	+3.35	+2.85
0.5 m		0 m	0	0.47	0.92	0.47	−72.00	−50.00	−27.00
			1	0.47	0.93	0.47	−74.00	−52.00	−26.00
			2	0.47	0.96	0.47	−80.00	−56.00	−31.00
1 to 3 m		0.5 m	0	0.40	0.95	0.68	−39.00	−26.00	−12.00
			1	0.20	0.96	0.85	−14.50	−9.00	−3.50
			2	0.05	0.98	0.97	−0.15	+0.35	+0.85

SIGNAL DETECTED WHEN THERE IS A HUMAN BODY BURIED IN A SNOW LAYER

The snowy ground is explored by transiting parallel to the antennas. The speed must be about 0.3 m/s. The response of the radiometer to movement over dry snow, about 1.6 m thick and with density about 0.38 g cm^{-3}, give fluctuations in the output potential lower than one volt. So the servo-system can keep the level of the output signal within reasonable limits, that is to say can define a local reference temperature.

Figure 5 represents the type of response of the radiometer to an empty hollow in the snow layer that has previously been defined. We obtain a signal with a maximum amplitude of about 4 V. Near a human body placed in the hole, the maximum amplitude of the output signal increases by 1.5 V or so.

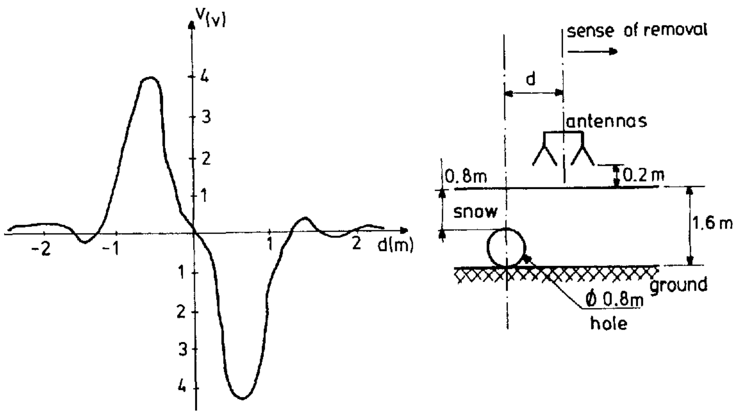


Fig. 5. Radiometer output signal when antennas move above a hole dug in a snow layer.

It should be noted that if the dry snow thickness changes a lot over distances of the order of one metre, fluctuations of the output potential increase and cause false alarms.

For a wet compact snow with a water content of around 2% in volume and a density of around 0.43 g cm^{-3}, the amplitude of the detected signals rapidly decreases. It is impossible to detect a human body in such snow except if its thickness is below 0.4 m.

False alarms on snowy ground

The directivity of the antennas used requires a systematic exploration of the snowy ground to be studied through parallel traverses separated by 2 m or so. Detection is made by an acoustic alarm which triggers from a 2 V signal. This is associated with an indicator like a galvanometer.

The ground studied had thickness irregularities of about 0.2 m. For dry snow 1 m thick, we obtained a false alarm every 20 m^2.

For a wet snow 1 m thick and 2% water content, there are practically no false alarms. In that case, we repeat that a victim can only be detected within 0.4 m of the surface, whereas in a dry snow, detection is certain up to 1 m, with an error rate less than 1%.

On an avalanche field, the systematic exploration by parallel traverses at 2 m intervals is impossible because of the irregular configuration of the snow layer.

Conclusion

The radiometric method is not a practical solution to the detection of avalanche victims, though contrary to the conclusion of Enander and Larson (1976), it seems possible to detect a human body under a snow layer if the victim is buried at a reasonable depth and if the water content of the snow is not too high.

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A RECORDING SNOW LYSIMETER

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ABSTRACT. An instrument which continuously records the run-off from a small, well-defined snow cover is described. The lysimeter is reasonably priced and effective, requiring only modest service attention; details of the records which it takes are given also. The relevance of the instrument to snow hydrology and to the energetics of a snow-cover is discussed.

RÉSUMÉ. *Un lysimètre à neige enregistreur.* On décrit un appareillage pour l'enregistrement continu des écoulements provenant de petites couches de neige bien individualisées. On donne une description d'un mécanisme efficace et de prix raisonnable pour un lysimètre à neige ne demandant qu'un entretien modeste, ainsi qu'un exemple des enregistrements obtenus. On discute ses possibilités d'emploi pour des études d'hydrologie nivale et spécialement du bilan énergétique du manteau neigeux.

ZUSAMMENFASSUNG. *Ein registrierendes Schneelysimeter.* Es wird eine Messanlage zur kontinuierlichen Aufzeichnung der Abflüsse aus kleinen, festumrissenen Schneedecken vorgestellt. Der Mechanismus eines kostensparenden, aber wirkungsvollen und zudem wartungsarmen Schneelysimeters wird beschrieben, seine Einsatztüchtigkeit durch ein Registrierbeispiel belegt. Ferner wird sein praktischer Nutzen bei schneehydrologischen und speziell schneedeckenenergetischen Fragestellungen diskutiert.

INTRODUCTION

Recent mass and energy balance studies of glaciers and snow fields of large area have revealed a need for basic investigations into the ablation mechanisms relating to ice and snow (Föhn, 1973; Wendler and Ishikawa, 1973; Ambach, 1976). A study into the development of temperate winter snow covers has been carried out by the Institut für Geographie der Universität München since 1971; this research is part of a long-term hydrological research programme (Herrmann and others, 1973). The present study investigates snow-cover development in a catchment region, of area 18.7 km², situated at the northern edge of the Bavarian Alps (670–1 801 m a.s.l.).

The lysimeter described here was developed in order to obtain detailed information about the run-off from alpine snow covers at low or medium altitudes. Such run-off is initiated by both melting and frequent winter rainfalls. Preliminary results from the study have already been published (Herrmann, 1974) and include estimates of ablation from the snow cover calculated from energy-balance data and from lysimeter records.

A simple water-sampling mechanism, involving only a few instrumental modifications, was added to the lysimeter in 1975 to permit the estimation of the isotopic content of the run-off.

INSTRUMENTATION SYSTEM

The lysimeter (Fig. 1) consists of two parts: a snow-collection area, and a recording unit, connected by a plastic tube.

The snow-collection area is square and 25 m² in area. This is probably the smallest size which can be used with any confidence that its results can be applied to larger snow covers. Larger catchment areas pose many research difficulties, in particular the problem of collecting and recording large quantities of water.

The snow catchment area is bordered by planks 0.1 m high. The ground within this border is covered with polyethylene sheet, 0.2 mm thick; this material is preferred to polyvinyl chloride as its thermal conductivity and radiative properties are more suitable. The ground of the catchment area slopes to a central outlet located in the middle of a border plank. We assume that the liquid which runs into the area from outside, at the upper edge of the area, is balanced by liquid which runs out at the lower border without being recorded. Funnel-shaped catchments ought not to be used, incidentally, as inflow which is not compensated by outflow can occur, with a circular design, and this leads to a systematic error. However, with a square area, liquid travel times must be taken into account in the calculations; water from different parts of the area will take different times to arrive at the outlet.

The water-collection system consists of a floatation chamber in the form of a plastic barrel (Fig. 2) installed in a pit. The barrel outlet, diameter 1.5 in (3.75 cm), is some distance above

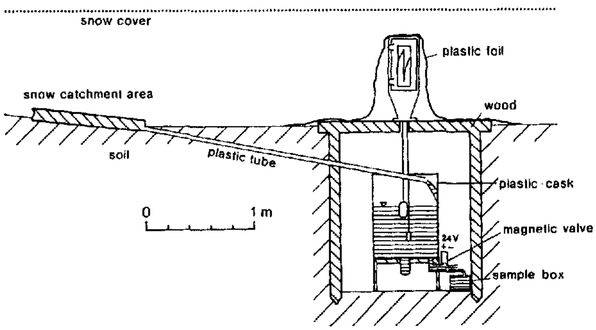

Fig. 1. The recording snow lysimeter.

Fig. 2. A photograph of the water-collection barrel and, above it, the water-level recorder.

the level of the pit base and is controlled by a 24 V d.c. magnetic valve. This allows the lysimeter to be controlled in remote areas where no mains electricity is available. Two standard accumulators (84 A h) are sufficient to supply the low current requirements of the valve for a whole winter.

The magnetic valve is controlled by two microswitches mounted in the recording unit of the lysimeter, which is an Ott vertical-drum water-level recorder with a variable recording scale (Fig. 2). These switches operate in the following way: When the water level reaches its maximum height of 0.25 m, the recorder pen closes the first microswitch. This allows current to flow to the valve, which opens, and allows water to run out of the barrel. As the water level drops, the recorder pen moves towards the minimum end of its scale eventually closing the second microswitch which shuts down the current and closes the valve again.

The recording drum of the water-level recorder is driven by a robust clockwork mechanism. The drum rotates once in 8 d producing a linear chart speed of 2 mm/h. Other speeds are possible as the transmission gears can be changed, but the speed quoted here was found to be the most suitable for establishing the relationships between the volume of water arriving from the catchment area and such quantities as cask dimensions and decay of the recording curves with time.

The pit which contains the recording and collecting units is covered by planks and snow. This maintains the pit at temperatures considerably above freezing point. However, additional protection is recommended for the recording unit which is made largely of metal. The formation of ice on the pulleys and belts of the transmission system can be minimized by isolating the recorder housing with "Styropor" plates. In an emergency, defrosting liquids can be sprayed into the unit to keep the recorder at work. A pit area of 1 m^2 is sufficient for the water from a previous emptying operation to have soaked into the ground before the next emptying, even with a pit bed of argillaceous morainic deposits.

The readings obtained using this system represent the mean water run-offs during the periods between emptying. If slight modifications of the basic unit are made (Fig. 3) then the water samples can be collected and retained instead of being lost to the ground. Such retention allows analysis of the snow-cover run-off, e.g. for its isotopic composition. The modification consists of the introduction of a central cavity into the bottom of the barrel in which the float is held, with the position of the barrel outlet below the level at which the second microswitch closes the outlet valve. Thus, nearly all the collected water leaves the barrel during the emptying operation. A certain proportion of this outflow can be lead to a 5 dm^3 sample box

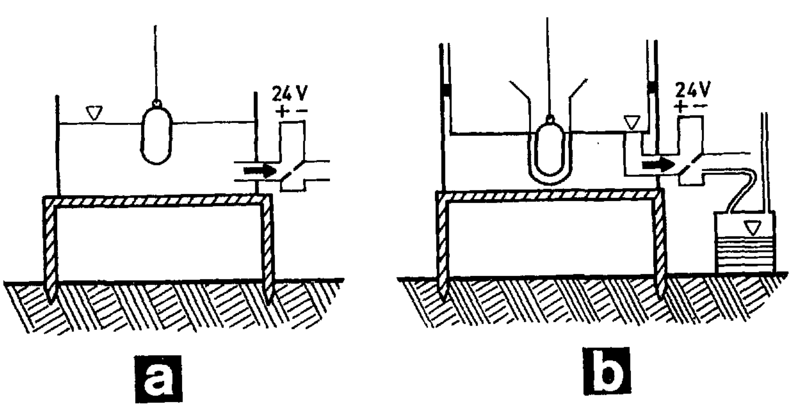

Fig. 3. Two alternative configurations for the lysimeter:
a. For the recording of run-off only.
b. For the recording and proportional retention of run-off.

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through a ¼ in (0.65 cm) diameter plastic tube inserted in the bottom of the magnetic-valve outlet. The proportion of liquid sampled is controlled using a clip on the tube. Thus, the isotopic proportions in a run-off system can be assessed by analysis of the sample collected.

Results

Figure 4 shows a recording produced by the snow lysimeter during a period of severe spring ablation. The chart speed was 2 mm/h and the recording scale 1 : 1.

The recording system described here draws curved plots while the floatation chamber is filling, and straight vertical lines while the chamber empties, a procedure which takes about 1 min. There is, at this chart speed, an upper limit for the accurate assessment of run-off. This upper limit corresponds to the frequent melting during day-time in the catchment area of 25–30 mm (water column) of the snow cover. For higher run-off rates the recording scale must be changed to 1 : 2.

An interruption of run-off is represented on the recording by a straight horizontal line. When this occurs at the top or bottom of the chart, it may also represent a valve switching failure.

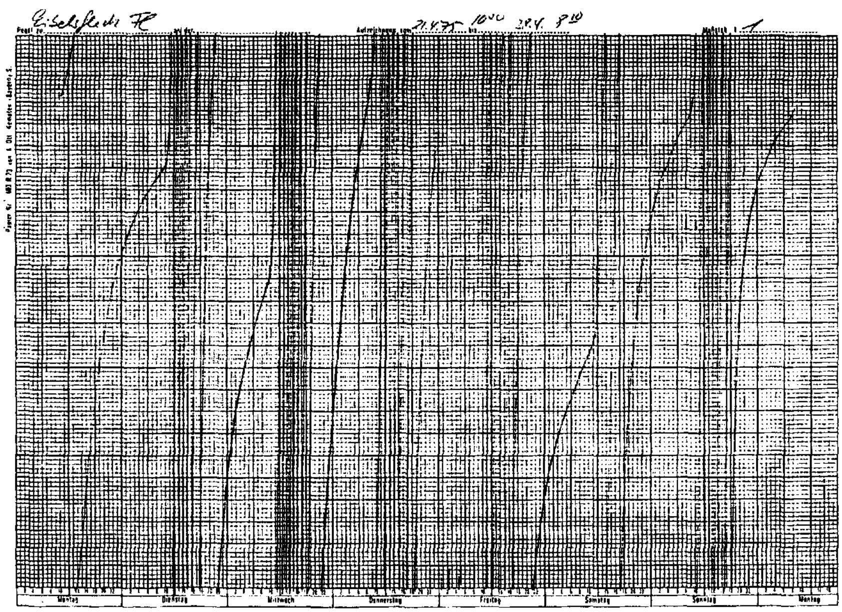

Fig. 4. A recording made by the lysimeter of the run-off from 25 m² of snow cover at 1 030 m a.s.l. in the Bavarian Alps, April 1975 (spring ablation). The 25 cm recording height corresponds to a collection of 75 l of liquid (c. *3 mm water column) when using a chart speed of 2 mm/h. (The recording which appears here has been photographically reduced.)*

Discussion

Two recording lysimeters of the type reported here are operating successfully in the research area in the Bavarian Alps, one in an open area and the other in a forest. The energy budgets of the snow catchment areas are estimated simultaneously from continuous readings of air temperature, snow temperature, humidity, wind speed, and radiation balance. An instrumental system which is sufficiently comprehensive can enable even very short-term studies to be made into such topics as the mass of local snow cover and variations in the energy balance (Herrmann, 1974).

Research of this kind yields data which may be used to formulate optimum models of the snow melt and simulations of run-off processes (involving both melting and rain) within a given catchment area (Herrmann, 1975). These simulations can be influenced to a considerable extent by the natural transmission and transformation properties of the areas studied, and these properties can be isolated and measured effectively by analysis of synoptic data from lysimeters and hydrographs. Furthermore, the encouraging preliminary experiments of Herrmann and Stichler (1976) into the isotopic content of snow have shown that the interpretation of reliable isotope data obtained with a lysimeter together with run-off data appears to allow some further quantitative insight into the relationships between the direct and delayed run-off of snow melt. Assessments of the contributions made by surface flow, interflow, and ground-water flow to total run-off during snow-melt periods can also be made.

ACKNOWLEDGEMENTS

I would like to extend especial thanks to Professor F. Wilhelm and Dr K. Priesmeier, both of the Institut für Geographie der Universität München for their co-operation during the planning and installation of the instrument described here. My thanks are also due to the Flussmeisterstelle Benediktbeuern/Oberbayern for carrying out the excavation work and to the Institut für Radiohydrometrie der Gesellschaft für Strahlen- und Umweltforschung mbH, München, who carried out the construction of the modified water-collection units for the snow lysimeters.

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SHORT NOTES

A SLIDE OF GLACIER ICE AND ROCKS IN WESTERN CANADA

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Abstract. A glacier-caused slide occurred in July 1975 within the Devastation Glacier basin, British Columbia. An estimated 2.5×10^6 m^3 of glacier ice, representing about 9% of the total mass, was incorporated into displaced debris. Four human lives were lost.

Résumé. *Une chute de glace de glacier et de rochers dans l'ouest du Canada.* Un glissement dû à un glacier s'est produit en juillet 1975 dans le bassin du Devastation Glacier en Colombie Britannique. Environ $2,5 \times 10^6$ m^3 de glace de glacier, représentant à peu près 9% de la masse totale, fut incorporée dans les matériaux déplacés. Quatre vies humaines furent perdues.

Zusammenfassung. *Eine Mure aus Gletschereis und Felsen in West-Kanada.* Im Becken des Devastation Glacier, Britisch-Columbia, wurde im Juli 1975 durch einen Gletscher eine Mure ausgelöst. Etwa $2,5 \times 10^6$ m^3 Gletschereis, das sind circa 9% der Gesamtmasse, wurden mit Schutt vermischt und verfrachtet. 4 Menschen kamen dabei ums Leben.

Location

A slide occurred in the upper basin of the Devastation Glacier (lat. 50° 36′ N., long. 123° 32′ W.) 1.2 km north-west of Pylon Peak (2 472 m) in Lillooet River watershed, Coast Mountain Range, British Columbia.

Description and causes of the slide

On 22 July 1975, a substantial mass of rocks, mud and ice, totalling an estimated $(27–30) \times 10^6$ m^3, broke away from the cirque of Devastation Glacier. It covered the entire length of Devastation Creek and blocked the confluence with Meager Creek. The slide originated under the ice of Devastation Glacier's accumulation area, at an elevation of 1 980 m, travelled 6.5 km and descended 1 150 m.

At present the active ice of Devastation Glacier is confined to the area above the cirque, about 3 800 m up-stream of where it was since the first record (Carter, 1932). The remnant of the tongue is a mass of stagnant ice covered by debris, still prominent in the narrow, sun-exposed valley, but has no visible connection with the active ice above.

From observations of the ice distribution over the catchment basin, it is concluded that the collapsed area in the slide was covered by glacier ice. As the slide went down, it gained momentum over at least one kilometre of dark-coloured stagnant ice in the valley and bounced off the valley walls at each sharp bend until it reached and dammed Meager Creek causing the formation of a small lake.

Geologically, the cirque area is underlain by Quaternary volcanic material, a part of Meager Creek volcanic complex (Read, 1977). At the base of the cirque, altered breccia, tuff, and ash deposits predominate. Poorly consolidated reddish-brown and grey volcanic rocks in the immediate area of the slide's origin form an unstable and apparently crumbly base beneath the ice cover (Fig. 1).

This poorly consolidated volcanic material incorporated in the main slide broke into comparatively large (2–3 m in diameter) blocks and boulders at the beginning of the descent. Further down, the slide material was broken into progressively smaller fragments. Within a few days, the debris at the terminus was converted into mud as it became saturated with water from Devastation Creek. Blocks of ice, over 10 m in size, mixed with the debris, were strewn over the entire path of the slide. The terminus of the deposited material was about 500 m wide and at the highest parts of the cross-section 30–40 m above the original ground level. This front stopped shortly after covering a sand bar near the confluence of Devastation and Meager Creeks. It was at this sand bar that the four men perished as they awaited the arrival of a helicopter.

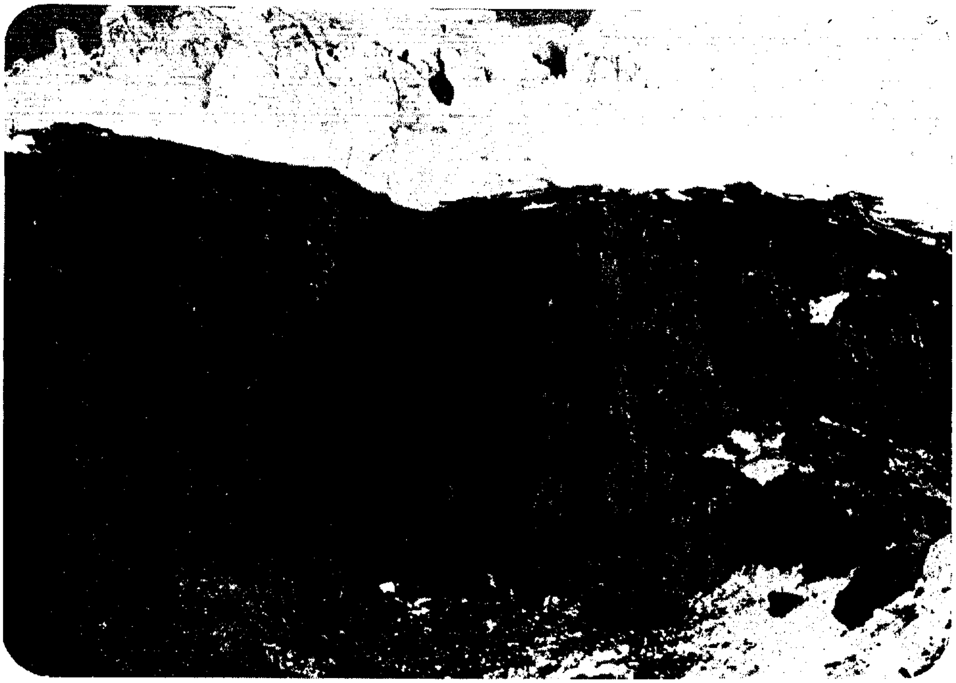

Fig. 1. Upper section of the slide area showing the crumbly base under the glacier ice. Dark vertical streaks are areas wet from melt water. Photograph 18 August 1975.

Fig. 2. Easterly part of the slide area. The ice thickness is about 48 m. Photograph 18 August 1975.

Entering Meager Creek Valley, the slide dammed the waters of Meager Creek, a tributary of Lillooet River, forming a lake and flooding the surrounding forest.

As no volcanic activities or tremors were recorded on that day for British Columbia (personal communication from R. J. Wetmiller), the slide is attributed to the first or both of the following causes:

1. The weight of the glacier ice and the action of glacier melt water. Melt water was discharging into the unstable ground material beneath the ice at an increasing rate because of favourable melt conditions. As the material became saturated with water, its decreasing bearing strength failed under the weight of the ice above and resulted in the collapse (Figs 1 and 2).
2. Some movement of ice in the form of a minor ice fall triggered the collapse of a large wet mass of supporting ground below the ice.

The volume of ice lost by the glacier is estimated at 2.5×10^6 m^3, approximately 2.5×10^6 tonnes or 9% of the estimated $(27\text{–}30) \times 10^6$ m^3 total volume of rock–ice slide.

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THOUGHTS ON OGIVE FORMATION

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ABSTRACT. The formation of ogives is examined from a structural point of view. It is suggested that the dark bands represent large-crystal, bubble-free ice which was transported from near the glacier base up toward the glacier surface by the processes of folding and/or reverse faulting due to compressive flow at the base of an ice fall, while the white bands represent relatively small-crystal, bubble-rich near-surface ice. The variation in compression to account for periodic folding and/or reverse faulting is explained by seasonal variation in flow velocity through an ice fall.

RÉSUMÉ. *Idées sur la formation des ogives.* La formation des ogives est examinée d'un point de vue structural. On suggère que les bandes sombres représentent des gros cristaux sans inclusions de bulles d'air, de glace qui a été transportée depuis les régions proches de la base du glacier jusqu'à la surface par le processus du plissement et/ou de failles de renversement dues à l'écoulement compressif à la base d'une chute de séracs, tandis que les bandes claires représentent des cristaux relativement petits riches en bulles de glace de surface. Les variations de compression dont il faut tenir compte pour expliquer la périodicité des plissements et/ou des failles de renversement sont expliquées par des variations saisonnières de la vitesse d'écoulement dans une chute de séracs.

ZUSAMMENFASSUNG. *Überlegungen zur Bildung von Ogiven.* Die Bildung von Ogiven wird von einem strukturellen Standpunkt aus betrachtet. Es wird angenommen, dass die dunklen Bänder aus grosskörnigem, blasenfreiem Eis bestehen, das aus der Umgebung des Gletscheruntergrundes zur Gletscheroberfläche durch Faltungs- und/oder Rückfaltungsprozesse unter dem Druckfliessen an der Basis eines Eisfalles transportiert wurde, während sich die weissen Bänder aus relativ feinkörnigem, blasenreichem, oberflächennahem Eis zusammensetzen. Die Druckschwankungen die zu periodischen Faltungs- und/oder Rückfaltungsvorgängen führen, werden aus den jahreszeitlichen Schwankungen der Fliessgeschwindigkeit in einem Eisfall erklärt.

OGIVES may be defined as transverse bands of alternating white and dark ice, convex down-glacier, emanating from the bases of ice falls on temperate glaciers. Several explanations have been suggested to account for this phenomenon (King and Lewis, 1961; Fisher, 1962). This study does not attempt to refute these theories but merely suggests alternative processes which could play a role in ogive formation.

King and Lewis (1961) suggest that the dark bands below an ice fall represent the glacier ice which descended the ice fall during summer. During this period ablation is maximum and accumulation minimum. Dirt and dust transported onto the glacier surface from nearby ice- and snow-free slopes, coupled with high ablation rate, result in surficial glacier ice which is characteristically comprised of mainly bubble-free, large ice crystals with high dirt content. In contrast, during winter when ablation is low and accumulation high and nearby slopes are mantled by fresh snow, ice crystals in glacier ice descending an ice fall are typically small with many bubble inclusions and little dirt content; thus forming the so-called white band. As King and Lewis (1961) correctly observe, these alternating bands may become accentuated down-glacier by differences in albedo between dark and light bands and concurrent accumulation of washed-in dirt in the resultant troughs.

The present study views this phenomenon from a structural glaciological point of view. It should be re-emphasized here that the writer does not disagree with the views put forth by King and Lewis (1961) but wishes to suggest a possible contemporaneous process. The velocity of glacier ice through an ice fall is greater than either immediately up-glacier or down-glacier, and, because discharge remains constant through this stretch, an increase in velocity must be associated with a decrease in cross-sectional area of the glacier through the ice fall. The shear stresses associated with the extending flow here are thought to initiate foliation parallel to the bedrock floor (Ragan, 1969). At the base of the ice fall an abrupt drop in valley gradient results in decreased velocity accompanied by compressive flow. Compression here results in folding and faulting of the folia (Ragan, 1969; Miller, unpublished). Folia attitudes below the Vaughan Lewis ice fall on the Juneau Icefield, Alaska, have been examined in detail by Miller (unpublished). He observes that most folia seem to dip down-glacier in the up-glacier sector of a white band and up-glacier in the down-glacier sector. While a wide range of dips is observed, "the highest percentages, however, center around 70° dip on each limb" of the white band. In addition Miller (unpublished) notes that the strike of these folia seems to parallel the form of the ogive. Ragan (1969) in mapping below ice falls observes that most folia seem to dip up-glacier.

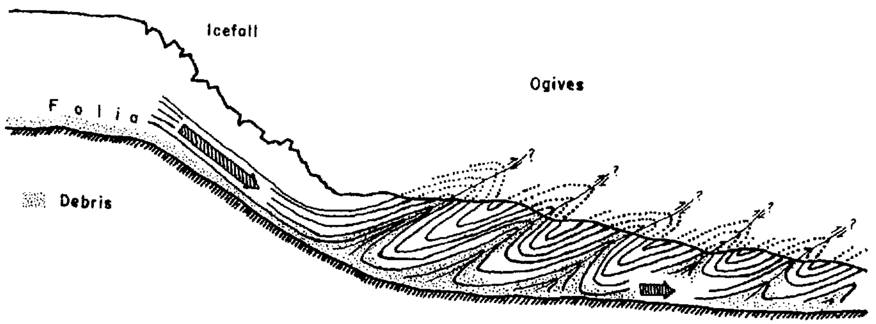

Fig. 1. Ogive formation according to folding with possible reverse faulting model.

Two possible interpretations are shown in Figures 1 and 2. Both interpretations are based largely on an assumption that the dark bands are not simply superficially dirt-covered as King and Lewis (1961) suggest. They noted that when a patch on a dark band which was cleared and swept of all dirt it appeared as white as the adjacent white band and that after one year again had become much dirtier than this white band. They conclude then that the dirt in the area of the dark band was "extremely sparse in the body of the glacier". There is, however, an alternative explanation. It would be enlightening to observe whether the cleared patch of dark band would have been as white as a *cleared* patch of white band. At this point on the glacier, some 800 m down-glacier from the ice fall, ablation should have resulted in a greater concentration of dirt on white bands than on similar white bands further up-glacier; thus a dark band swept clean might well be as white as a relatively dirty white band. The dirtying process on the dark band would then quickly reclaim the cleared patch by the processes of ablation and redistribution of dirt by supraglacial melt-water streams. Nevertheless, if the dark bands should prove to be truly dirt-free at depth the blue bubble-free nature of this ice could still indicate deep-glacier origins for this ice. The models illustrated in Figures 1 and 2 would then have to be slightly altered to indicate that basal dirt does not reach the glacier surface though the mechanism of folding and/or faulting could still occur.

Figure 1 illustrates generalized isoclinal folding below an ice fall, as suggested by Ragan's (1969) data. This folding may or may not be associated with axial plane reverse faulting. Miller's (unpublished) data would indicate symmetrical rather than isoclinal folds and again with possible axial plane faulting. Such tight folding could well bring those folia which were formerly in close proximity to the bedrock floor, high up through the glacier. Axial plane faulting would further enhance this situation. These folia with their dirt-rich ice may then be exposed by subsequent ablation. The light-colored, bubbly nature of the white bands, on the other hand, is typical of ice which had stayed close to the glacier surface at all times. Due to differential ablation the result is thus a classic case of inversion of topography with synclinal ridges and anticlinal troughs as ablation proceeds slowly on the high-albedo white bands and more rapidly on the low-albedo dark bands.

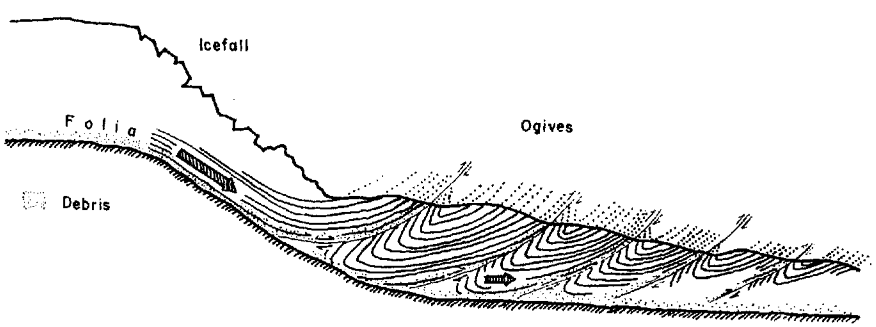

Fig. 2. Ogive formation according to reverse faulting with associated drag folding model.

Figure 2 illustrates a possible alternative structural explanation. Here rather than folding being the predominate mechanism, reverse faulting with associated drag folding is emphasized. Again the folia attitude shown is that conforming generally to Ragan's (1969) observations. It would not be difficult, however, to illustrate the same mechanism so that it would conform to Miller's (unpublished) observations. Here, reverse faulting due to compression at the base of an ice fall is responsible for lifting dirt-rich, bottom-most tectonic folia toward the glacier surface. Again, these dirty zones become the dark bands and subsequent troughs down-glacier from the ice fall. Even if the dirt-rich folia should not reach the surface, these zones would still be comprised of ice typically bubble-free and follow the same mechanism as outlined above.

The seeming annual nature of ogives can be explained readily to be compatible with either of the aforementioned models. Because data regarding seasonal fluctuation of glacier velocity through an ice fall unfortunately is sparse and inconclusive, until further data become available the following is suggested as a reasonable tentative hypothesis. It is suggested here that a decrease in glacier velocity through an ice fall in winter and an increase in summer could account for the annual formation of ogives. In summer months increased amounts of water acting as an englacial and sub-glacial lubricant could result in a velocity increase over this steep stretch where basal slip is of such importance to glacier flow, while during winter months a decrease in supraglacial, englacial, and subglacial melt water through this attenuated stretch of the glacier could result in a velocity decrease. Increased summer velocity would result in increased compression at the ice fall base with corresponding reverse faulting and/or anticlinal folding in this area. Lower winter velocity, on the other hand would result in decreased compression here with corresponding synclinal folding and/or absence of reverse faulting.

Much detailed field work, including structural mapping, year-round surveying, and bore-hole deformation studies are of course essential in order to confirm or reject these suggested models. Nevertheless, it is hoped that these models will stimulate further examination of this unique glacial phenomenon.

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