THE REMAINS OF ROCK GLACIERS IN BOTTELNEK, EAST CAPE DRAKENSBERG, SOUTH AFRICA

By

COLIN A. LEWIS\(^1\) AND PATRICIA M. HANVEY\(^2\)

\(^1\) Department of Geography, Rhodes University, Grahamstown 6140, South Africa
\(^2\) Department of Geography and Environmental Studies
University of the Witwatersrand, Johannesburg 2050, South Africa

SUMMARY

Certain debris accumulations in Bottelnek are ascribed a rock glacier origin on the basis of morphological and sedimentological evidence. Radiocarbon dating indicates that rock glaciers were active at or subsequent to 21 000 BP and that cold, stadial conditions, existed in the region after 27 000 BP and before 13 000 BP, during the Bottelnek Stadial. At least sporadic permafrost existed in Bottelnek when the rock glaciers were active.

INTRODUCTION

The morphology and sedimentology of unconsolidated slope deposits in Bottelnek (31° 06'S; 27° 45'E) that resemble rock glaciers, is examined in order to understand their origin and to help resolve debate as to whether or not permafrost existed in the region during the Quaternary. Sediments deposited under periglacial conditions have already been recorded in Bottelnek (Hanvey & Lewis 1991) and elsewhere in the vicinity (Lewis & Dardis 1985, Hanvey \textit{et al.} 1986, Lewis & Hanvey 1988, 1991). In reviews of periglacial studies in southern Africa Lewis (1988a, b, 1990a) suggests that periglacial features are widespread in the uplands of Lesotho and adjoining areas of the Drakensberg. Active periglacial features have been noted at altitudes above 1 900 m above sea level, while inactive features are more widespread. Fitzpatrick (1978) postulated that permafrost formerly occurred in the uplands of the Drakensberg. Active rock glaciers have not been recorded in southern Africa, although Marker (1990, 1992) attributes scree tongues in the Amatola Mountains and 'in the southern African high country' to a rock glacier origin.

REGIONAL SETTING

Bottelnek is an east-west aligned valley in the uplands of the Drakensberg approximately mid-way between Elliot and Barkly East in Cape Province (Fig.1). The floor of the valley, which is at altitudes of between 2 000 and 1 820 m, is incised some 400 m below the summit plateau of the Drakensberg. The valley is eroded into Clarens sandstone overlain by basaltic rock of the Drakensberg Formation (Beukes 1970, South African Committee for Stratigraphy 1980). In cross profile the valley is asymmetric, with steep basaltic cliffs on its northern flank and more gentle slopes,
Fig. 1. Debris accumulations, some of which resemble rock glaciers, on the northern flank of Bottelnek.
predominantly developed on Clarens sandstone, but normally culminating in minor cliffs cut in basalt, on the southern flank. Superficial deposits exposed in sections up to 30 m deep occur on the northern side of the valley. They form both terrace and tongue-like features on the side and floor of the valley. Much of the southern side of Bottelnek consists of bare outcrops of Clarens sandstone or of sandstone overlain by a thin veneer of colluvium. The valley is floored with terraced alluvial sediments. The present study is concerned with terrace and tongue-like features on the northern side of the valley.

MORPHOLOGY OF THE TONGUE-LIKE AND TERRACE FEATURES

Twenty eight accumulations of superficial debris, some of which resemble rock glaciers, have been identified on the northern side of Bottelnek (Fig. 1). Many of these are elongate tongue-like features, some of which extend out from small hollows within the basalt (e.g. Numbers 1, 2, 7 and 9). Others show a much more equable length : width ratio, forming broad terrace-like features towards the valley bottom (e.g. Numbers 3, 5, 6 and 8). In eight consecutive cases (Fig. 1, Numbers 11–18) these features are separated from the basaltic cliffs above them by a region in which no accumulations are discernible, except for individual basaltic clasts which litter the surface. The eight features terminate abruptly, either in debris faces that are presently being eroded by the Bottelnekspuit, or in longitudinally convex fronts that rest on Clarens sandstone bedrock (e.g. Fig. 1, Number 16). A series of shallow channels cut into bedrock lead downslope from the terminus of feature 16 (Fig. 1) and end in rounded moulin-like holes that descend toward the present channel of the Bottelnekspuit.

The tongue-like features identified vary in length from approximately 500 m to 250 m. In cross-profile they are invariably convex. These features vary in width between approximately 250 m and 30 m. The tongues normally begin at the foot of gulleys that extend down cliff faces at small valley heads or on the flanks of the main valley side, although in some cases, as already noted, the tongues are detached from cliffs and couloirs. The longitudinal surface of the tongue-like features is usually inclined at angles of less than 20°, although 30° slopes exist on the upper segments of some tongues (e.g. Number 21). Some of the tongue-like features are demarcated by linear depressions and levees aligned parallel, or sub-parallel to the long axis of the features. The fronts of the tongue-like features are often characterized by a number of bulges or lobes, as at Rose Hill (Number 2), while others have a unified terminus. The height of the front of tongues as measured from ground level at the base of the front, varies from 4 m to 30 m. Contact between tongue debris and bedrock is, in many cases, not exposed.

Terrace-like features are up to approximately 200 m long and vary in width between approximately 250 m and 200 m. The fronts of the terraces vary in height between 20 m and 25 m. The surface of these features undulates, although the amplitude of the undulations is within a range of 5 m. The gradient of the surface slope of these terraces immediately above the front usually varies between 10° and 15°, but steepens up slope towards the contact with the rock outcrops from which the terrace debris appears to originate. Clasts are scattered on the surface of terraces.
STRATIGRAPHY OF THE TONGUE-LIKE AND TERRACE FEATURES

The stratigraphy of the features being examined is well exposed by fluvial erosion caused by the Bottelnkpspruit. The tongue-like forms exhibit a repetitive pattern with a coarser facies overlying a finer facies. Both facies consist of basalt and sandstone clasts held within a finer sandy matrix. A water reworked layer was identified in places between the upper and lower facies. Occasional water reworked layers also occur within the lower facies. The deposits exposed in the fronts of the Rose Hill and Chesney Wold tongues (Numbers 2 and 21, Fig. 1), which are discussed in detail later in this paper, epitomize the tongue deposits.

Slope wash deposits appear more common in the terrace features than in the tongues. Two organic layers, the lower of which resembles a palaeosol, exist in feature 5 (Fig. 1). The upper layer, which is 2 cm thick, lies 2 m above river level and is overlain by 12 m of horizontally bedded colluvium. This upper layer has been radiocarbon dated to 21 000 ± 400 BP (Pta — 5654).

DETAILED STUDIES

Two of the tongue-like features were examined in detail in terms of morphology, stratigraphy, sedimentology, surface morphology of quartz grains, and clast characteristics. These were Rose Hill (Number 2, Fig. 1) and Chesney Wold (Number 21, Fig. 1).

Rose Hill

Rose Hill is a tongue of unconsolidated but largely vegetated debris (Figs 2 and 3) which begins at the base of a series of major gulleys within a small valley head on the northern flank of the Bottelnkpspruit (Fig. 1). The tongue extends from the base of the exposed bedrock, at an altitude of 2 060 m, to terminate on the floor of Bottelnk at 1 930 m a.s.l., where it is partially undercut by the Bottelnkpspruit. The tongue is approximately 400 m long and varies in width from ± 70 m (cross-section 2, Fig. 2) to almost 160 m near the toe of the tongue (cross-section 5, Fig. 2). The morphology of the tongue is shown on Fig. 2. Except at its western extremity, where the terminus of the tongue is actively eroded by the Bottelnkpspruit, a river terrace divides the tongue from this stream.

The stratigraphy and sedimentological characteristics of the Rose Hill tongue are apparent only at its lower extremities, where fluvial erosion has exposed sections up to 12 m high (Fig. 4). Bed rock consisting of Clarens sandstone, is overlain by four sedimentary units: a matrix supported diamicton (Unit One) succeeded in ascending order by stratified granule layers with basaltic cobbles (Unit Two), a fine facies composed mainly of sand and silt with angular clasts (Unit Three), and by a coarse facies in which clasts are clast to matrix supported and up to 2 m in length, with an inclusion of stratified pebbly gravels and granules (Unit Four; Fig. 4).

Particle size analysis of the fine fraction, using the wet sieving and pipette method, as outlined by Ingram (1971) and Galehouse (1971), was undertaken on one sample from each of the four sedimentary units. All samples consist predominantly of sand
Fig. 2. The morphology of the Rose Hill fossil rock glacier.
(74.3%–59.3%), with subordinate quantities of silt (31.6%–20.2%) and a lesser amount of clay (9.4%–5.6%; Fig. 5). The uppermost sedimentary unit (Unit Four) contains more sand than any other unit. All samples display bimodality, the more marked peak being the very fine sand fraction (+3−+4 φ), with a secondary peak in the very coarse sand fraction (−1−0 φ).

Ten quartz grains from the fine sand fraction (+2−+3 φ) were selected from samples from each of the sedimentary units for surface texture analysis using the electron microscope. The grains were prepared for examination using the method recommended by Krinsley & Doornkamp (1973). The gross morphology of the grains (cf. Powers 1982) reflects the sedimentary units from which they were derived (Table 1).

Table 1

<table>
<thead>
<tr>
<th>Unit</th>
<th>Subrounded</th>
<th>Characteristics of grains (%)</th>
<th>Subangular</th>
<th>Angular</th>
</tr>
</thead>
<tbody>
<tr>
<td>Four</td>
<td>60</td>
<td>40</td>
<td></td>
<td>−</td>
</tr>
<tr>
<td>Three</td>
<td>20</td>
<td>60</td>
<td>20</td>
<td></td>
</tr>
<tr>
<td>Two</td>
<td>60</td>
<td>40</td>
<td></td>
<td>−</td>
</tr>
<tr>
<td>One</td>
<td>20</td>
<td>50</td>
<td>30</td>
<td></td>
</tr>
</tbody>
</table>
Fig. 4. The stratigraphy of sediments exposed at the lower end of the Rose Hill rock glacier tongue.
Fig. 5. Particle size characteristics of the fine fractions of sedimentary Units exposed at the lower ends of the Rose Hill and Chesney Wold debris accumulations.
Table 2 records the surface features noted on those quartz grains which displayed features diagnostic of glacial (frost shattered) origins.

Table 2

Surface features diagnostic of glacial (frost shattered) origins recorded from quartz grains collected at Rose Hill

<table>
<thead>
<tr>
<th>SURFACE FEATURES</th>
<th>Conchoidal fracture</th>
<th>Linear/sub-linear fracture</th>
<th>Parallel and sub-parallel steps</th>
<th>Adhering particles</th>
<th>Dish-shaped concavity</th>
<th>Striae</th>
<th>Blocky surface</th>
<th>Plyy surface</th>
<th>Silica overgrowth</th>
</tr>
</thead>
<tbody>
<tr>
<td>Number of features</td>
<td>10</td>
<td>5</td>
<td>8</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>1</td>
</tr>
<tr>
<td>Four</td>
<td>4</td>
<td>8</td>
<td>4</td>
<td>7</td>
<td>2</td>
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<tr>
<td>Three</td>
<td>10</td>
<td>5</td>
<td>4</td>
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<td>2</td>
<td>5</td>
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<td>18</td>
<td>14</td>
<td>2</td>
<td>1</td>
<td>2</td>
<td>2</td>
<td>1</td>
</tr>
<tr>
<td>Total</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Features as a % of total features recorded (Total = 96)

| 33 | 25 | 18,8 | 14,6 | 2,1 | 1 | 2,1 | 2,1 | 1 |

Features as a % of total grains examined (Total = 40)

| 80 | 60 | 45   | 35   | 5   | 2,5 | 5   | 5   | 2,5 |

Fig. 6 depicts examples of a variety of surface features existent on quartz grains sampled from Rose Hill. These include conchoidal fractures (A, C); linear and sub-linear fractures (B); parallel and sub-parallel steps (C, base of photo); linear and sub-linear fractures and adhering particles (D); adhering particles (E); and irregular blocky surfaces (F). These latter surfaces (F) occur on the majority of quartz grains scanned from Rose Hill.

Clast characteristics were recorded at seven sites on the Rose Hill tongue (Fig. 2). Fifty clasts were sampled at all sites and their roundness was determined by visual comparison with assessment images (based upon Krumbein 1941). The length of the long axis of each clast was also measured.

Fig. 7 depicts size and sorting variability within and between samples. Individual clasts range in length from 5 to 250 cm. Median values range between 24 and 70 cm. Sorting, as indicated by the interquartile range, is best developed in sample 1, which records clasts from Unit Three as exposed in the front of the tongue (Fig. 2). The least well sorted sample occurs on the surface at the western edge of the tongue (sample 3), where a wide range of large clasts occurs. Regression analysis (Fig. 7 inset), shows that median clast size on the surface of the tongue varies with distance along the tongue, with
Fig. 6. Quartz sand surface textures, Rose Hill. A: conchoidal fracture; B: linear/sub-linear fractures; C: conchoidal fracture, parallel steps (base of photo); D: linear fractures, adhering particles; E: detail of adhering particles; F: irregular blocky surfaces and depressions. A, B, C, D, F are from Unit Four; E is from Unit Three.
Fig. 7. Size and sorting of clasts on and within the Rose Hill rock glacier deposits. The location of the sample sites is shown on Fig. 2.
clasts becoming smaller towards the terminus or distal region. Positive correlation between surface clast size and distance exists at the 99% level.

Clast roundness at Rose Hill (Fig. 8) varies between Units Three and Four exposed in the front of the tongue and with distance along the tongue (Fig. 8, inset). Sample 5, which may include fluviolally transported debris (being adjacent to the ravine that parallels the eastern side of the tongue) as well as clasts that have moved down tongue by non-fluvial processes, has a smaller percentage of angular clasts than its position on the lobe might suggest (Table 3). Regression analysis (Fig. 8) shows that clast roundness varies with distance along the tongue, although positive correlation only exists at the 58% level.

Table 3
Clast angularity on the surface of the Rose Hill tongue *

<table>
<thead>
<tr>
<th>Sample number: (Ranked in relation to distance down tongue)</th>
<th>% of Angular Clasts (Classes 1–4)</th>
<th>% of Rounded Clasts (Classes 5 and upwards)</th>
</tr>
</thead>
<tbody>
<tr>
<td>6</td>
<td>50</td>
<td>50</td>
</tr>
<tr>
<td>5</td>
<td>26</td>
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<td>7</td>
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<td>62</td>
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<td>4</td>
<td>30</td>
<td>70</td>
</tr>
<tr>
<td>3</td>
<td>20</td>
<td>80</td>
</tr>
</tbody>
</table>

*Samples 1 and 2 were based on clasts sampled in section at the end (front) of the tongue.

Chesney Wold

This is a tongue of unconsolidated debris that originates at and beside a basalt outcrop at an altitude of 2135 m on the north side of Bottelnek, and terminates in an 18 m high bluff that is presently being eroded by the Bottelnekspruit on the floor of the valley at 1840 m (Fig. 1, Number 21). The tongue is approximately 550 m long and varies in width from 75 m to 100 m over most of its length, but widens to 190 m where it terminates on the floor of Bottelnek. This tongue shows distinct curvature along its long axis (Fig. 9). A shallow depression runs on either side of the tongue and, in cross-section, the tongue is generally convex. The amplitude of the relief between the crest of the tongue and the adjacent parallel depressions varies from 5 to 10 m. A number of discontinuous ridges of rocky debris run longitudinally along the tongue, although they are obscured by shrubby vegetation in places as are minor depressions and lobate features (Fig. 9).

Rock is exposed in two cliffs close to the head of the tongue. The easterly cliff is composed of amygdaloidal basalt whereas the westerly cliff is of denser basalt. Joint blocks in the westerly cliff average 0.9 m × 0.67 m in length and breadth, whereas in the eastern cliff they average 0.3 m × 0.41 m. The debris tongue originates at the base of the western cliff.

The stratigraphy and sedimentological characteristics of the Chesney Wold tongue are evident only where the tongue has been eroded by the Bottelnekspruit on the floor of the main valley. Bedrock, of Clarens sandstone, is overlain by four sedimentary units (Fig. 10).
Fig. 8. Clast roundness on and within the Rose Hill rock glacier deposits. The location of the sample sites is shown on Fig. 2.
Fig. 9. The morphology of the Chesney Wold fossil rock glacier.
Fig. 10. The stratigraphy of sediments exposed at the lower end of the Chesney Wold rock glacier tongue.
Particle size analysis of the fine fraction, using the same method as at Rose Hill, was undertaken on one sample from each of the four sedimentary units. All samples consist predominantly of sand (83–60.9%), with subordinate quantities of silt (33.6–14.5%) and a lesser amount of clay (12–2.5%; Fig. 5). The lowest sedimentary unit (Unit One) contains more sand than any other unit. All samples display bimodality, the more marked peak being the very fine sand fraction (+ 3 – + 4 φ).

Quartz sand grains were selected for surface texture analysis, as at Rose Hill, from each sedimentary unit and studied using the electron microscope. Irregular blocky surfaces, similar to that depicted on Fig. 6 (F), were common. Conchoidal fractures, linear and sub-linear fractures, parallel and subparallel steps, and adhering grains were also identified.

Clast characteristics, using the same sample size and criteria as at Rose Hill, were recorded at six sites on the Chesney Wold tongue. Individual clasts range in length from 4 to 230 cm. Median values range between 10 and 48 cm (Fig. 11). Sorting, as indicated by the interquartile range, is best developed in sample 1, recording clasts from Unit Two as exposed in the front of the tongue (Fig. 9). Median clast size varies with distance along the tongue, with clasts becoming smaller towards the terminus of the tongue. Positive correlation between surface clast size and distance exists at the 86% level.

Clast roundness (Fig. 12) exhibits a reduction in angularity with distance down slope, with positive correlation between angularity and distance existing at the 99% level.

INTERPRETATION

(i) Morphology

The morphology of both the Rose Hill and Chesney Wold debris tongues (Figs 1, 2 and 9) is similar to that of many fossil tongue rock glaciers (Domaradzki 1951, Wahrhaftig & Cox 1959, White 1976) as described from Alaska (e.g. Calkin et al. 1987, Capps 1910, Wahrhaftig & Cox 1959), the Antarctic (e.g. Hassinger & Mayewski 1983), Argentina (e.g. Barsch 1988, Corte 1976, 1987); Austria (e.g. De Jong & Kwadijk 1988), Switzerland (e.g. Barsch 1977, 1978; Haeberli 1985), France and Italy (e.g. Evin 1987), Central Asia (Gorbunov & Titkov 1989), Xinjiang, China (Francou et al. 1990), the western U.S.A. (e.g. Morris 1987, White 1976) and other parts of the world (Barsch 1988, Vitek & Giardino 1987). Linear ridges and depressions are characteristic especially of the lower portions of tongue rock glaciers, while in many cases the tongues are fed by debris that has moved down gulleys that seam the cliffs above the rock glaciers (Wahrhaftig & Cox 1959, Embleton & King 1975) and/or down scree slopes (Haeberli 1985, Wahrhaftig & Cox 1959). Tongue rock glaciers are also commonly convex in cross-section (e.g. Barsch 1988, Capps 1910, Wahrhaftig & Cox 1959, Haeberli 1990).

Wahrhaftig & Cox (1959), in their study of rock glaciers in the Alaska Range, state that many rock glaciers 'are separated from their source cliffs either by a small glacier . . . or by a . . . pit . . . that marks the position of a small recently melted glacier'. White (1976) has made similar observations. The latter feature characterizes debris tongues 11
to 18 (Fig. 1) in particular, which are separated from the cliffs above them by a boulder strewn area in which debris tongues are not apparent and may be the remains of ice cored rock glaciers.

The distribution along the flanks of the Bottleneck valley of broad terrace-like deposits, (e.g. Fig. 1, Number 5) and their undulating relief are both characteristics of lobate rock glaciers (Wahrhaftig & Cox 1959, White 1976, Shakesby et al. 1987).

(ii) Stratigraphy

The stratigraphy of features studied in Bottleneck, of coarse material overlying finer material, is characteristic of rock glaciers. Wahrhaftig & Cox (1959) record 'coarse rubble mixed with silt, sand, and fine gravel' beneath an upper coarse rubble unit. Embleton & King (1975) support this observation, as does Corte (1976). Haeberli (1985)

Fig. 11. Size and sorting of clasts on and within the Chesney Wold rock glacier deposits. The location of the sample sites is shown on Fig. 9.
Fig. 12. Clast roundness on and within the Chesney Wold rock glacier deposits. The location of the sample sites is shown on Fig. 9.
shows that in active rock glaciers fine material is largely absent in the coarse surface layer, because that is the active layer above the permafrost table, but that fine material is 'abundant underneath the permafrost table'. Sediments exposed at the front of the Rose Hill and Chesney Wold debris tongues, as well as in other debris tongues in Bottelnek, accord with the above descriptions of rock glacier sediments. The granule and boulder facies that form Unit Two at Rose Hill and Unit One at Chesney Wold and fine upwards into sand at Chesney Wold, and which include imbricated rounded basalt clasts, appear to be fluvial deposits. Rock glaciers commonly descend to valley floors and overlie, and sometime interdigitate with, fluvial deposits (Gorbunov 1988, Gorbunov & Titkov 1989).

(iii) *Particle Size*

Evin (1987, 1988) has shown that the lithology of the rocks from which rock glaciers are derived is important in deciding their internal composition, but that sand and silt is the dominant texture of the fine matrix with clay being rare or absent. Evin (1987) attributes these proportions of sand and silt to frost weathering and further suggests that clay which had been formerly present in rock glaciers may have been removed, in part or in whole, by the melting of interstitial ice. The particle size distribution presented by Evin (1987, 1988) compares favourably with the results from Rose Hill and Chesney Wold (Fig. 5).

Wilson (1990a, b) has also shown that sand is the dominant constituent of the fine matrix of fossil rock glaciers in northwest Ireland. Although the clay percentages at Rose Hill and Chesney Wold are higher than those recorded by Wilson (1990a, b), the overall predominance of sand and silt supports the interpretation of the fine matrix material as being rock glacial in origin.

(iv) *Clast Characteristics*

Wahrhaftig & Cox (1959) state that 'average fragment size varies considerably from place to place on a single rock glacier', but that longitudinal banding of debris is common. They traced the bands to the different areas of the backwall from which the debris is derived and found that joint patterns on the backwalls influence the size of debris particles. Fragment size also varies considerably on the Rose Hill and Chesney Wold tongues (cf Fig. 7), and at Chesney Wold the debris tongue originates at the base of a cliff that consists of larger joint blocks than adjacent cliffs. This supports the interpretation of the debris tongues as rock glacial.

On both the Rose Hill and Chesney Wold debris tongues the median size of clasts on the surface of the tongues decreases with distance down tongue. Yarnal (1982), as well as Gorbunov & Titkov (1989), show that clasts tend to increase in size with distance down rock glaciers. At least some clasts on the surface of the Rose Hill and Chesney Wold tongues may postdate deposition of the main debris accumulations. Surface clast analyses at Rose Hill and Chesney Wold are therefore unlikely to elucidate the origins of the bulk of the debris accumulations at either site.

(v) *Surface Morphology of Quartz Grains*

Although there have been few studies of the surface characteristics of quartz grains from rock glaciers, Corte & Trombotto (1984) have reported that 'The grinding action of
the rock glacier . . . produces arched steps, concoidal (sic) fractures . . . similar to . . .

Ying Wang et al. (1982) list five features that they consider to be glacial
textures: conchoidal fractures, parallel steps, striae, upturned plates, dish-shaped
concavities. Trewin (1988), based on the work of Higgs (1979), lists seven features as
abundant on quartz grains that have been subjected to glacial processes: conchoidal
fractures, straight and arcuate steps, fractures, adhering particles, straight scratches,
angular outline, low relief; as well as three features common on quartz grains that have
experienced glaciation: parallel striations, imbricated grinding features, curved scratches.

Surface features identified on quartz grains from Rose Hill and Chesney Wold
suggest that these grains have been subjected to glaciation (cf. Culver et al. 1983).
Nevertheless the grains selected for study were those that exhibited such characteristics,
whereas the majority of grains displayed other, particularly blocky features. In view of
Wilson's (1990b) conclusion when considering a similar predominance of blocky
features and conchoidal fractures, the evidence from Bottleneck is not conclusive of a
glacial origin, although it indicates frost shattering and movement that caused
conchoidal fracturing, linear fractures and the other features. Further examination of
quartz grain textures from active rock glacier environments is needed to verify this
conclusion.

(vi) Conclusion

The evidence of morphology, stratigraphy, particle size of the matrix, the
characteristics of clasts within the debris tongues, and surface characteristics of quartz
grains suggests that the Rose Hill and Chesney Wold tongues are fossil rock glaciers.
Evidence from the back wall of the Chesney Wold tongue, which originates under a
jointed cliff producing large blocky clasts, is also significant. Wahrhaftig & Cox (1959)
note that 'Rock glaciers occur on blocky fracturing rocks which form talus that has large
interconnected voids on which ice can accumulate', and similar observations have been
made by other scientists (e.g. Parson 1987). The overall conclusion is that many of the
debris accumulations present along the northern flank of the Bottleneck Valley are fossil
rock glaciers.

RADIOCARBON DATING

Organic material from debris accumulation 5 in Bottleneck (Fig. 1), which is believed
to be a lobate rock glacier, dates to 21 000 ± 400 BP (Pta — 5654), indicating that the
rock glacier was existent at or subsequent to that date. This accords with evidence of cold
climatic conditions at 20 900 + 350 BP (Pta — 4944) reported by Opperman &
Heydenrych (1990) from Strathalan Cave B near Maclear, which is ± 60 km east of
Bottleneck. This cave was occupied intermittently by Middle Stone Age people from an
unestablished date until at least 22 500 ± 230 BP (Pta — 4859) but was abandoned
before 20 900 BP. Subsequent re-occupation of the adjacent Strathalan Cave A did not
take place for 'at least 10 000 years' (Opperman & Heydenrych 1990). At Sehonghong in
Lesotho a microlake industry bears witness to human occupation of that site at 13 000
± 140 BP (Pta — 884), although previous occupation apparently terminated around
19 860 + 220 BP (Pta — 918), probably due to climatic and associated conditions that rendered the area unsuitable for human occupation (Carter & Vogel 1977).

Lewis & Dardis (1985) report head deposits at Dynevors Park in the valley to which Bottelnek is tributary. These deposits overlie organic material, containing insect remains, that date to 31 600 ± 950 BP (Pta — 5657). The insect remains, which have not yet been fully analysed (M. Hill, pers comm), may indicate interstadi al conditions. Some 28 km northeast of Bottelnek, at Birnam, organic-rich layers that accumulated in a lake (Hanvey & Lewis 1990), range in age from 35 000 ± 1 500 BP (Pta — 5263) at the base of the lake sediments to 27 700 ± 1 000 BP (Pta — 5246) at the top of those sediments.

Angular rock fragments, possibly deposited as a result of periglacial processes, intrude into the lacustrine sediments at Birnam. Organic material close to the base of these fragments dates to 24 300 ± 370 BP (Pta — 5088). The evidence suggests that subsequent to 24 300 BP cold, periglacial conditions succeeded less rigorous conditions in which the lake existed.

At Stillorgan, in the Riflespruit valley to which the Birnam valley is a tributary, organic-rich material dating to 27 000 ± 510 BP (Pta — 5260) is overlain by what may be slope or glacial deposits (Hanvey 1990) supporting the conclusion that harshening of climate took place subsequent to that date.

The evidence from the East Cape Drakensberg and Lesotho suggests that interstadi al conditions that existed by 35 000 BP were replaced by the onset of colder, stadial conditions subsequent to 27 000 BP; that by 21 000 BP rock glaciers existed in Bottelnek and that human occupation of the area (vide Hanvey & Lewis 1991) apparently ceased; but that warmer interstadi al conditions suitable for human occupation existed by 13 000 BP. The stadial in which rock glaciers existed in Bottelnek is hereby named the Bottelnek Stadial and is essentially of the same age as the Dimlington Stadial in England (26 000 to 13 000 BP; Rose 1985); a cold phase in Australia (29 000 to 13 000 BP; Chappell 1991); the Otira Glaciation in New Zealand (‘somewhat before 22.3 ka BP’ to ‘ca. 14 ka. BP’; Suggate 1990); and the Llanquihue moraines of Patagonia (‘more recent than 30 000 BP, older than 14 000 BP’; Rabassa & Clapperton 1990, but see Broecker & Denton 1989); but is apparently younger than the Kalambo Interstadi al of East Africa (c. 35 000 to 25 000 BP; Mahaney 1990, Coetze 1967, van Zinderen Bakker & Coetze 1988) and the Tullabardine Interstadi al of Tasmania (ca. 50 000 to 25 000 BP; Colhoun & Fitzsimmons 1990), which may equate with the supposed interstadi al insect-containing deposits that underlie head at Dynevors Park.

PALAEOClimatic SIGNIFICANCE:

Many scientists (e.g. Barsch 1977, 1978, Haeberli 1973, 1975, 1985, King 1976, 1982, 1985) believe that active rock glaciers are indicators of at least discontinuous permafrost. Whalley & Martin (1992) argue that they can also be found in other climatic areas although King et al. (1992) imply that only ‘inexperienced geomorphologists’ come to that conclusion and that active rock glaciers are ‘the most important indicators of mountain permafrost’. Haeberli (1985) writes that ‘Rock glaciers are permafrost phenomenon of cold, dry mountain regions’ that only develop below the equilibrium line on glaciers and above the
lower permafrost limit. The lower altitudinal boundary of discontinuous permafrost equates with a mean annual air temperature of $-2^\circ\text{C}$, although the temperature necessary for rock glacier formation is partly dependent upon mean annual precipitation and may be much lower than $-2^\circ\text{C}$ (Haebler 1985).

Assuming that at least discontinuous permafrost was necessary for rock glacier formation in Bottleneck, the fossil rock glaciers in that valley, which probably originated subsequent to 27000 BP and became inactive prior to 13 000 BP, indicate that mean annual air temperatures were $-2^\circ\text{C}$, or lower, during at least part of that time. This is at least $14^\circ\text{C}$ lower than the mean annual air temperature recorded at Barkly East, the nearest meteorological station to Bottleneck, at present (Lewis 1990b). Permafrost was present when the rock glaciers were active.

Gorbunov (1979) introduced the concept of specific density of rock glaciers, which Titkov (1988) describes as 'the ratio of the total area of rock glaciers in the individual basin (sq. m) to the area of the basin above the isohypse of the lower limit of occurrence of rock glaciers (sq. km)'. Titkov (1988) states that the greatest specific density of active rock glaciers occurs when the degree of glaciation is between 10 and 20%. Below 10% either the altitude (and hence the mean annual temperature) or the precipitation is too low for rock glacier development. The presence of fossil rock glaciers in Bottleneck therefore suggests that true glaciers may have existed synchronously in that vicinity. Because rock glaciers occur below the equilibrium line on glaciers (Haebler 1985), the features in Bottleneck suggest that the equilibrium line on such glaciers and ice caps as existed in the adjacent Drakensberg when the rock glaciers were active, lay above 2600 m, which is the upper altitude of the back walls from which debris was supplied to the rock glaciers of Bottleneck. Finally, the rock glaciers indicate that arid or semi-arid conditions existed in Bottleneck (and by implication the remainder of the East Cape Drakensberg) during at least part of the time span between 27000 and 13000 BP.

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REFERENCES


