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The origin of slope deposits in the southern Drakensberg, eastern Lesotho

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CHAPTER 9

Discussion

9.1 INTRODUCTION

This chapter will analyse the results for the deposits, in terms of whether or not they comprise of characteristics typical to the process origins described in Chapter 2. The process origins will be discussed in terms of the hypotheses set out in Chapter 1, therefore debris flows, landslides, mudflows/mudslides, solifluction deposits and glacial deposits will be discussed for all sites. Fluvial deposits will only be referred to in the context of the Sehonghong sites, whilst rock glaciers and pronival ramparts will only be discussed in terms of the linear slope deposits. The morphological and sedimentological characteristics of the Drakensberg deposits will be compared to those for the various process mechanisms as described in Chapter 2. The feasibility of the proposed process origins will then be discussed for each individual deposit, based on the analysis of the results, in order to determine which processes formed the various Drakensberg features.

9.2 MORPHOLOGY

9.2.1 Debris flows/avalanches

The morphology of the Tsatsa-La-Mangaung, Sekhokong Site 1 and Leqooa Valley deposits do not resemble that of a debris flow, given that these types of deposits are known to develop distinctive features as discussed in section 2.1.2.2.1 (Figures 2.3, 5.2, 6.2 and 7.3). In contrast, although the Sekhokong Site 2 deposit also lacks the presence of an erosional scar and lateral levees, waves are apparent in plan view and in the longitudinal cross section (Figure 9.1), which may indicate a debris flow origin. The presence of bedrock exposures however, may have diverted flow, suggesting that the waves are not necessarily indicative of a debris flow. An example of a debris flow in the Drakensberg, measuring approximately 230 m in length and 135 m in width at its frontal lobe, is described by Grab (1999b) (Figure 9.2). Flow levees are also present at this site (Grab, 1999b); however, levees or lobate snouts are absent at the linear deposits described in this PhD project, which suggests that the Tsatsa-La-Mangaung, Sekhokong Site 1 and Leqooa Valley deposits are not debris flows.

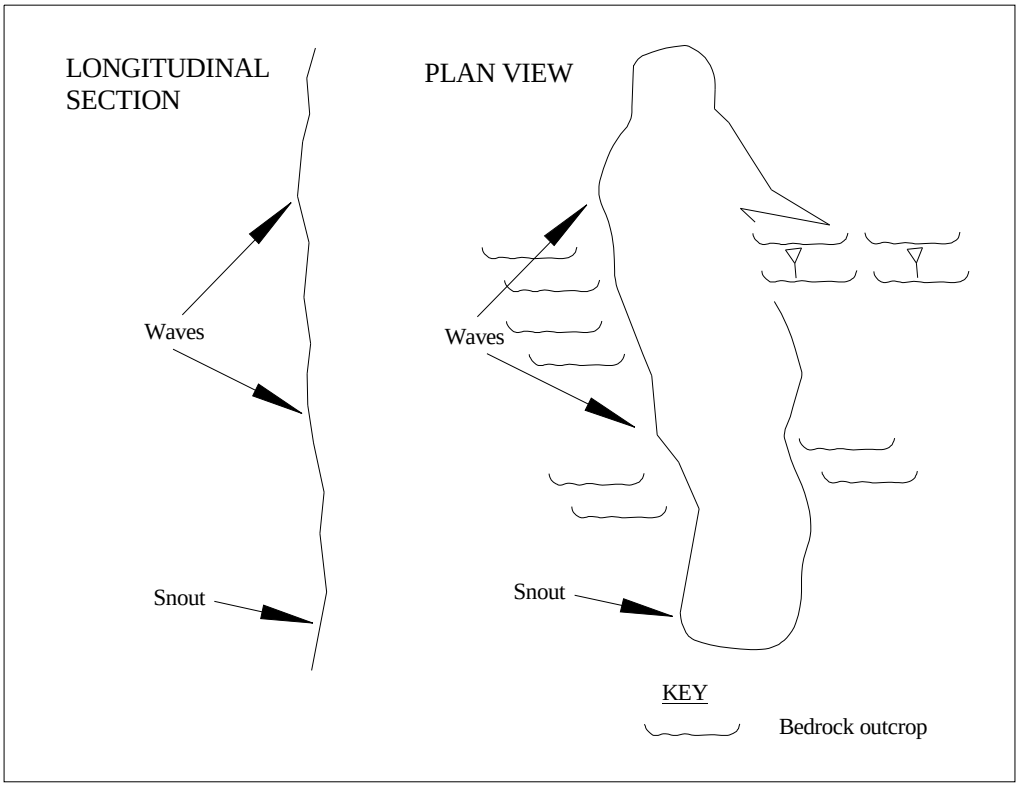


Figure 9.1 A plan view and a longitudinal cross section of the Sekhokong Site 2 deposit showing waves.



Figure 9.2 Debris deposits in the Drakensberg at Njesuthi (photo by S. Grab).

The topographic setting at all linear deposit sites may be conducive to debris flow activity. The slope angle above the Tsatsa-La-Mangaung deposit ranges between 29° and 45° (Figure 5.2; cross section E-E), whilst the slope angle above the Sekhokong Site 1 deposit ranges between 27º and 36º (Figure 6.2; cross section E-E). In addition, the slope angle above the Sekhokong Site 2 deposit is 26º (Figure 6.20; cross section E-E), whereas it ranges between 18º and 35º above the Leqooa western deposit and 18º and 29º above the eastern deposit (Figure 7.3; cross sections I-I and J-J). These slope angles all fall within the range of those from which debris flows become initiated as described in section 2.1.2.2.1. In contrast, the slope angle above the Sehonghong south-facing deposits is a maximum of 15°, whilst that above the north-facing deposits reaches a maximum of 17° (Figure 8.2; cross section D-D and E-E), which reduces the likelihood of debris flow activity. There are no areas of colluvial infills above any of the deposits, neither does there appear to be a sufficient source area of material, which is problematic when considering a debris flow/avalanche origin. However, in mountainous regions, snowmelt within small drainage basins with steep slopes may encourage debris flow processes (Selby, 1993).

The triangular plan shape of debris avalanches as described in section 2.1.2.2.2 is not observed in the morphology of the deposits. However, the angle above the Tsatsa-La-Mangaung, Sekhokong and Leqooa Valley deposits is similar to threshold slope angles described for debris avalanches (Figures 5.2, 6.2, 6.20 and 7.3). The slope angle above the Sehonghong deposits reaches a maximum of 15º on the south-facing slope and 17º on the north-facing slope (Figure 8.2), suggesting that these slope angles would not be steep enough for the mobilization of debris avalanches.

9.2.2 Landslides

There is an absence of a former erosion scar above all deposits, which is typical of landslides. However, the scar may have degraded since the deposits formed and may no longer be recognisable (Wilson, 2004). The lack of a failure scar should therefore not preclude the consideration of slope failure mechanisms in the formation of the deposits (Wilson, 2004). The length-width ratio for the various deposits ranges between 7:1 and 16:1 at the Tsatsa-La-Mangaung slope deposit, between 4:1 and 10:1 at the Sekhokong Site 1 deposit, and between 5:1 and 21:1 at the Leqooa western deposit, which falls within the range of length-width ratios of landslides as described

in section 2.1.2.2.3. Alternatively, the length-width ratio of the Sekhokong Site 2 deposit ranges between 7.5:1 and 3.5:1, whilst it ranges between 4:1 and 7:1 for the Leqooa eastern deposit, which does not fall within the typical range of landslides. The depth/length (D/L) ratio of the Tsatsa-La-Mangaung deposit is 7% at its greatest depth, whilst it is 8% at both Sekhokong deposits, which place them in the debris slides and avalanche category. In contrast, the D/L ratio of the Leqooa Valley deposits is 4%, which is just below the threshold of D/L ratios for debris slides and avalanches (Table 2.5).

The theory on hillslope orientation for landslide occurrence on equator-facing slopes does not concur with the Tsatsa-La-Mangaung, Sekhokong, Leqooa Valley and Sehonghong south-facing deposits, given that these occur on south-facing slopes, which are pole-facing. Alternatively, the Sehonghong north-facing deposits are equator-facing and concur with the occurrence of landslides on these slopes. In contrast, the occurrence of landslides on south-facing slopes in the Western Cape as described in section 2.1.2.2.3 is consistent with the Tsatsa-La-Mangaung, Sekhokong, Leqooa Valley and the Sehonghong south-facing deposits, however all these deposits lack the typical morphological attributes of landslides.

9.2.3 Mudflows/mudslides

The Drakensberg deposits lack the typical morphological attributes of mudflows and mudslides described in section 2.1.2.2.4 (Figure 5.1, 6.1, 6.19, 7.1 and 7.2). The slope angle of the Tsatsa-La-Mangaung deposit is 25° at the head of the deposit, and levels off to 15º towards the toe of the deposit (Figure 5.2; cross section A-A), which is similar to the slope angle of the mudflow described by Corominas (1995) and falls within the range of typical mudflow slope angles (Brunsden, 1984). The Sekhokong Site 1 deposit tends to be fairly uniform (22° to 20°) (Figure 6.2; cross section A-A), whilst the Sekhokong Site 2 deposit has an average slope gradient of 22º, with steeper portions up to 36° (Figure 6.20; cross section A-A), all of which fall within the range of mudflows. In addition, the average slope angles at the Leqooa Valley deposits are 11º for the western deposit and 12º for the eastern deposit (Figure 7.3; cross sections A-A and E-E), which again fall within the range of gradients for mudflows and mudslides. In contrast, the clear lack of morphological characteristics at all sites, suggest that these deposits are unlikely to be of a mudflow or mudslide origin.

9.2.4 Solifluction lobes

The linear deposits attain significantly greater heights and lengths than are typical of solifluction deposits as described in section 2.1.2.2.6. The linear deposits also vary considerably in terms of their morphology in comparison to solifluction lobes in the Drakensberg (Figure 9.3) and they occur on upper slope positions (Figures 5.2, 6.2, 6.20 and 7.3), which is not the preferred location for solifluction deposits. In addition, the slope above the Tsatsa-La-Mangaung and Sekhokong deposits exceeds 25° (Figure 5.2, 6.2 and 6.20), which indicates that solifluction lobes would have been likely to collapse on such steep slopes. In contrast, the slope above the Leqooa Valley deposits ranges between 18° and 22° (Figure 7.3), which suggests that solifluction lobes would have survived. Stone-banked solifluction lobes have been identified in the Drakensberg on slope gradients between 14º to 19º (Grab, 2000b), which is again much gentler than the slope gradients recorded for the linear deposits.



Figure 9.3 Solifluction lobes on the Drakensberg (photo by S. Grab).

The median length/tread width form ratio of solifluction lobes has been described as 0.7 (Hugenholtz and Lewkowicz, 2002). The median length/tread width ratio is 7.5 at Tsatsa-La-Mangaung, 3.6 at Sekhokong Site 1, 3.5 at Sekhokong Site 2, 4.6 at the

Leqooa western deposit and 4.1 at the Leqooa eastern deposit, all of which are considerably larger than the length/tread width of solifluction lobes.

Although the Sehonghong deposits do not resemble solifluction lobes, they may however be a product of solifluction processes. Solifluction and gelifluction may occur on slopes of less than 5º and those of over 15º (French, 1996). This concurs with the slope angles on both the north- and south-facing slopes of the Sehohong River Valley (Figure 8.2). The slope gradient ranges between 1º and 15º on the south-facing slope (cross section D-D) and 1º and 17º on the north-facing slope (cross section E-E). Solifluction also causes valley sides to become convexo-concave in profile (Ballantyne and Harris, 1994) and mass wasting is believed to flatten and smooth the landscape (Ballantyne and Harris, 1994; French, 1996), which is observed on both the north- and south-facing slopes (Figure 8.2).

9.2.5 Rock glaciers

Although the linear deposits fall within the classification of tongue-shaped rock glaciers described in section 2.1.3 and display similar dimensions, there is a distinct lack of the typical attributes of a lobate form, or ridges and furrows at the surface (Figures 5.1, 6.1, 6.19, 7.1 and 7.2). Although slight ridges and furrows are observed at Sekhokong Site 2 (Figure 6.20; cross section A-A), relict rock glaciers are usually recognised in the field by their furrows and ridges, which are subdued, however still prominent (Barsch, 1996), which is not the case at the Sekhokong Site 2 deposit. According to Barsch (1996), relict rock glaciers should still be 5 m to 10 m thick at the front ridge, however the Sekhokong Site 2 deposit is only 3 m high at its front (Figure 6.19 and 6.20). The absence of distinct ridges and furrows suggests that this deposit is unlikely to be of rock glacier origin.

9.2.6 Pronival ramparts

Although the linear deposits fall within the size dimensions of recorded pronival ramparts, the deposits occur linearly downslope and not parallel to the mountain slope. The Tsatsa-La-Mangaung and the Leqooa eastern deposit have short proximal slopes and long distal slopes (Figure 5.2; cross section C-C, Figure 7.3; cross-sections F-F to H-H), which concurs with those for pronival ramparts as discussed in section 2.1.4. In contrast, the Sekhokong Site 1 deposit has a long proximal slope and a short

distal slope as indicated by cross-sections B-B and C-C (Figure 6.2), which does not concur with those of pronival ramparts. The Sekhokong Site 2 deposit has a longer eastern slope (Figure 6.20; cross sections B-B to D-D), which would correspond to pronival ramparts if this was the distal slope, however the topographic setting does not allow for the identification of whether the slope would be proximal or distal to a snow patch (Figure 6.20).

The slope angle at the Tsatsa-La-Mangaung, Sekhokong Site 1 and Leqooa Valley deposits is significantly lower than the observed maximum distal slope angles of pronival ramparts, however this may be a result of slumping over time. The Sekhokong Site 2 deposit has an eastern slope angle of 35°, which corresponds with maximum distal slope angles for pronival ramparts, however its topographic setting does not allow for the differentiation between proximal and distal slopes. In contrast, suggestions that distal and proximal slope angles should be symmetrical and vary in terms of their cross sectional form (Behre, 1933; Harris, 1986; Shakesby *et al.,* 1995) is reflected in the Tsatsa-La-Mangaung deposit where cross sections A-A and C-C (Figure 5.2) are almost symmetrical, with the exception of cross section B-B which has a steeper distal zone. The Leqooa western slope deposit tends to have symmetrical proximal and distal slopes as indicated by cross-sections B-B to D-D (Figure 7.3), which concurs with observations by Behre (1933).

9.2.7 Glacial moraines

The Tsatsa-La-Mangaung and Leqooa eastern deposits have gentle proximal and steep distal slopes (Figure 5.2 and 7.3), which is similar to those observed for push moraines (section 2.1.5.1). However, push moraines are usually less than 10 m in height (Benn and Evans, 1998), which is considerably less than the recorded height for all linear deposits. Alternatively, the linear deposits fall within the size range of lateral moraines. The Sekhokong Site 1 deposit is the only one to show evidence of oversteepening on its proximal side (Figure 6.2), which has been described from a lateral moraine in Norway (Vere and Matthews, 1985), however the lower gradients and the lack of proximal slope oversteepening at the other sites could be attributed to slumping over time.

The secondary ridges observed at the Tsatsa-La-Mangaung, Sekhokong Site 1 and Leqooa Valley sites may indicate the extent of a former glacier (Figures 5.2, 6.2 and 7.3). Moraines may be deposited successively during progressive glacier recession and mark the extent of the former maximum extent of a glacial advance (Knight, 2004). Although these secondary ridges are smaller than the main ridge, which does not really reflect the advance of a glacier, this may however reflect the supply of material at the time of formation. The lack of gullying and dissection of these secondary ridges indicates that they are not a result of slumping of material from the main ridge and may represent glacier retreat.

9.2.8 Fluvial deposits

This section only applies to the Sehonghong deposits as the linear slope deposits were clearly not deposited under fluvial conditions. The Sehonghong deposits attain similar heights to depositional terraces described in section 2.1.6. The Sehonghong south-facing deposits also follow the river for approximately 2 km and the north-facing deposits follow the river for a few hundred metres (Figure 8.2), which is similar to river terraces. Parts of the same river terrace may become separated by tributary entrenchment or other geomorphic processes and therefore may be discontinuous along a valley (Harden, 2004), as indicated in Figure 9.4. Relict fluvial terraces can reflect the accumulation of alluvial sediment before subsequent downcutting (Lowe and Walker, 1997), which is believed to be the most appropriate theory for the Sehonghong deposits, given the thickness of material which has accumulated.

9.2.9 Summary

The most likely process origin for the Tsatsa-La-Mangaung, Sekhokong Site 1 and Leqooa valley deposits based on their morphology alone is that of a glacial moraine. The Sekhokong Site 2 deposit does not possess all the typical attributes of any process origin, however based on morphology it is possible that this deposit may reflect debris flow processes. The morphology of the Sehonghong deposits resembles that of a fluvial terrace; however it is also possible that solifluction processes contributed to the accumulation of material.

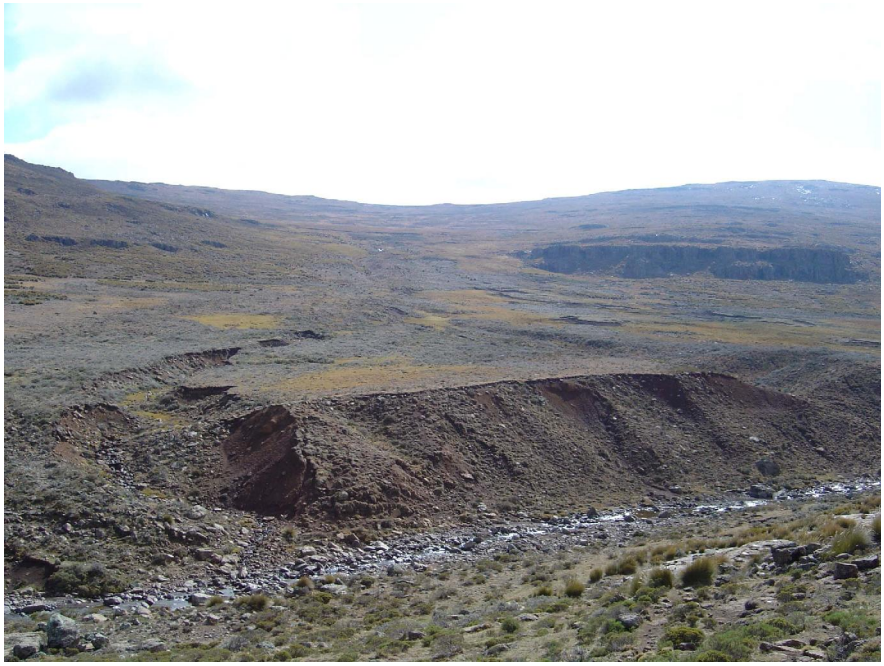


Figure 9.4 An example of where a potential terrace has been separated by tributary entrenchment on the south-facing slope at Sehonghong (photo by S. Mills).

9.3 PARTICLE SIZE ANALYSIS

9.3.1 Debris flows/avalanches

The linear slope deposits, the Sehonghong south-facing deposits and the upper unit of the Sehonghong north-facing deposit consist of similar particle characteristics to debris flows (section 2.1.2.2.1). All deposits mainly contain <4% clay (Tables 5.1, 6.1, 6.12, 7.1, 7.4 and 8.1), which is typical of debris flows. Debris flow diamicts and gravels have been described as typically bimodal (Scott, 1971; Phillips and Davies, 1991), and mesokurtic (Owen, 1994). The Tsatsa-La-Mangaung deposit displays a unimodal trend (Table 5.1), which is not characteristic of debris flows, however several of the particle size distributions within the Sekhokong Site 1 deposit are bimodal in trenches 1 and 3 (Table 6.1). The Sekhokong Site 2 deposit is also bimodal in trench 2 at 80 cm and 120 cm depth (Table 6.12) and the particle size characteristics at Leqooa Valley display several bimodal trends in trenches 1 and 3 of the western slope deposit and in trench 1 of the eastern slope deposit (Table 7.1 and 7.4). In addition, the sediment at Sehonghong Site 1 at 60 cm depth and site 2 at 120 cm depth is bimodal and the sediment at 45 cm depth is mesokurtic (Tables 8.1 and 8.2), which also concur with debris flow characteristics. The sediment within the exposure on the north-facing slope consists of coarse clasts within a fine-grained

matrix, which is poorly sorted and bimodal at 150 cm and 260 cm depth (Table 8.1), and reflect similar attributes to those of debris flow characteristics. The Sekhokong Site 2 deposit contains coarser material in trench 3, which would represent the snout of the deposit and which concurs with the coarse material within debris flows dominating flow heads (Takahashi, 1978, 1991; Johnson and Rodine, 1984).

9.3.2 Landslides and mudflows/mudslides

The clay content within the Tsatsa-La-Mangaung and Sekhokong Site 1 deposits is <3%, with the exception of Sekhokong Site 1 in trench 3, where clay content is <7% (Tables 5.1 and 6.1), whilst clay content within the Sekhokong Site 2, Leqooa Valley and Sehonghong deposits is generally <2% (Tables 6.12, 7.1, 7.4 and 8.1). This is significantly lower than the typical clay content within mudslides and landslides and there is no evidence of increased clay content with depth. Whilst clay fractions may have been eroded out through exfiltration processes, one would still expect a greater clay fraction at depth, if it were of a mudflow/mudslide origin.

9.3.3 Solifluction lobes

The linear slope deposits particle size characteristics do not concur with typical attributes of solifluction lobes as described in section 2.1.2.2.6 (Tables 5.1, 6.1, 6.12, 7.1 and 7.4). Solifluction landforms typically involve a diamict consisting of fines and large particles up to about 10 cm to 20 cm. Although the linear deposits also involve a diamict, this unit is considerably larger (over 1 m in depth) (Figures 5.4 to 5.6; 6.3 to 6.5; 6.21 to 6.23; 7.4 to 7.9) and clasts excavated from within the deposits were also over 1 m in length, which is considerably larger than those found within solifluction deposits. Solifluction deposits also tend to contain low amounts of clay-sized particles (Elliot and Worsley, 1999), which concurs with the low clay content observed within the linear slope deposits (Tables 5.1, 6.1, 6.12, 7.1 and 7.4). It has also been suggested that silt and clay particles tend to be concentrated near the fronts of lobes (Benedict, 1970; Hugenholtz and Lewkowicz, 2002), however this is not observed within the linear slope deposits, in fact at Sekhokong Site 2 the opposite is true and coarse material has accumulated near the front of the deposit (Table 6.12). None of the deposits contain buried organic soil horizons, which occurs in almost all solifluction lobes (Benedict, 1976; Kinnard and Lewkowicz, 2006), suggesting that the linear slope deposits do not owe their origins to solifluction processes.

Although the Sehonghong deposits do not resemble solifluction lobes, they may nevertheless be a product of solifluction and gelifluction, associated with mass wasting. According to French (1996), gelifluction deposits include coarse talus, alluvial silt and clay and diamicts of glacial or periglacial origin, which are usually matrix-supported, with the matrix commonly consisting of clay, silt or silty sand. This is true of all units in the exposures, where the upper unit is predominantly coarse, whereas the lower units in the south-facing exposure are predominantly made up of sand and the lower units on the north-facing exposure are predominantly made up of silt (Table 8.1). Head deposits are believed to be a product of solifluction and tend to be poorly sorted and comprise of matrix supported diamicts (Ballantyne and Harris, 1994; French, 1996), which concurs with the sediments within the south-facing exposures and the sediment within the upper units of the north-facing exposure (Table 8.2).

9.3.4 Rock glaciers

The linear deposits do not possess the typical two layered system of relict rock glaciers described in section 2.1.2.2.6. A relict rock glacier described by Humlum (1998a) contains similar sized rock fragments to those excavated from the Tsatsa-La-Mangaung and Sekhokong Site 2 deposit, however there is a distinct lack of finer material present in all trenches at Tsatsa-La-Mangaung and in trenches 1 and 3 at Sekhokong Site 2 (Tables 5.1 and 6.12). The Sekhokong Site 1 and Leqooa deposits on the other hand contain significant quantities of fine material (Tables 6.1, 7.1 and 7.4), but lack the two-layered structure of rock glaciers. The sediment within all linear deposits tends to be poorly sorted (Tables 5.2, 6.2, 6.13, 7.2 and 7.5), which agrees with rock glacier sorting. The sediment is also platykurtic in trenches 1 and 3 of the Sekhokong Site 1 deposit (Table 6.1), trench 2 of the Sekhokong Site 2 deposit (Table 6.13) and in the Leqooa western and eastern slope deposits (Tables 7.2 and 7.5), which have also been described as rock glacier attributes (Vere and Matthews, 1985).

9.3.5 Pronival ramparts

The sediment characteristics of the Tsatsa-La-Mangaung and Sekhokong Site 2 deposits (Tables 5.1 and 6.12) are similar to those described in section 2.1.4 for pronival ramparts, which consist of coarse, clastic, poorly sorted sediments, with an absence of fines even at depth (Washburn, 1979; Linder and Marks, 1985; Ballantyne

and Kirkbride, 1986; Hall and Meiklejohn, 1997). In contrast, the Sekhokong Site 1 and Leqooa Valley deposits contain a significant amount of fine-grained material (Tables 6.1, 7.1 and 7.4), which has also been described from pronival ramparts (Shakesby *et al.*, 1995).

9.3.6 Glacial moraines

The Tsatsa-La-Mangaung and Sekhokong Site 2 deposits possess the characteristics of supraglacial material described in section 2.1.5.2, consisting of coarse, unimodal sediment (Tables 5.1 and 6.12). In contrast, the Sekhokong Site 1 and Leqooa Valley deposits display several bimodal distributions and the presence of finer material (Tables 6.1, 7.1 and 7.4), which indicates that active processes may also have taken place at these sites. In terms of sorting, tills host a wider range of particle sizes than most other sediments, thus making them extremely poorly sorted (Gale and Hoare, 1991; Owen, 1994), which concurs with the poorly sorted material within the linear slope deposits (Tables 5.2, 6.2, 6.13, 7.2 and 7.5).

Bennett and Glasser (1996) describe flow tills as being poorly consolidated, which concurs with particle packing within the Sekhokong Site 2 deposit, which is primarily composed of unconsolidated material (Figures 6.21 to 6.23). Individual flow packages may also be visible in flow tills and sorted sand and silt layers, associated with reworking by meltwater are common (Bennett and Glasser, 1996). Successive flows of debris may also create upper washed horizons and interbeds of silt, sand or gravel (Benn and Evans, 1998). This may be observed within trench 3 (Figure 6.23), where a sandy unit is present at 130 cm depth, however this unit is very poorly sorted, as opposed to the well sorted layers described by Bennett and Glasser (1996).

The Sehonghong deposits are not regarded as moraines due to their morphology, however the fact that they may comprise till cannot be ruled out. Tills tend to be poorly sorted and lack stratification (Summerfield, 1991), which is observed in the sediment on the south-facing exposure (Figures 8.3 and 8.4), however deposits modified by meltwater may exhibit stratification, which is also observed in the sediment on the north-facing exposure (Figure 8.5). The coarse upper layer on the south-facing slope is predominantly composed of gravel and contains <11% fines (Table 8.1), which could indicate clasts that have been transported supraglacially.

Alternatively, the upper gravelly unit on the north-facing slope contains between 19% and 40% fines and is bimodal, which may indicate a certain degree of active transport. The lower sedimentary units on the south-facing exposure contain much finer particle sizes and are also bimodal, with the exception at 220 cm depth at Site 1, and may also infer a greater degree of active transport than in the upper unit. The compact massive sediment in the lower units of the Sehonghong Site 3 deposit is also similar to the structure of tills, as is the sharp non-gradational boundary between till units (Benn and Evans, 1998), which also occurs within the sedimentary units at this site (Figure 8.5).

9.3.7 Fluvial deposits

Site 1 and Site 2 along the south-facing exposure of the Sehonghong deposits contain >32% gravel in the lower unit and >54% gravel in the upper unit (Table 8.1), which suggests that if these were fluvially deposited sediments, they would have been deposited under a high energy environment, which is unlikely given that the length of the Sehonghong River and all its tributaries is only 9 km (Grab *et al.*, in prep). The upper units along the north-facing exposure are also unlikely to be of fluvial origin given the size of boulders present (Figure 8.5), which would have once again required a very high energy environment for deposition. These may however have been deposited by catastrophic flooding (Grab *et al.*, in prep). On the other hand, the lower units consist of >56% silt, with sand making up the remaining proportions (Table 8.1), which is similar to overbank deposits. These overbank sediments are normally stratified gravels, sands, silts and clays (Bridge, 2004). Stratification is not present within the lower units of the south-facing deposits (Figures 8.3 and 8.4); however alternating layers of sand and gravel occur within the north-facing deposit (Figure 8.5), which concurs with fluvial sediments.

9.3.8 Summary

All the linear deposits display attributes similar to the nature of debris flow deposits and pronival ramparts. The linear deposits also display the characteristics of till, however the particle size characteristics of the Tsatsa-La-Mangaung, Sekhokong Site 2 and Leqooa eastern deposits are similar to supraglacial debris, whereas the Sekhokong Site 1 and Leqooa Valley deposits are similar to debris which has undergone active transport. The particle size analysis of the Sehonghong deposits indicates that solifluction and gelifluction processes may have contributed to the

formation of these deposits. In addition, both the south- and north-facing deposits display attributes similar to till, whilst the north-facing deposits may also be a result of fluvial deposition based on particle size analysis results.

9.4 CLAST FABRIC AND FABRIC EIGENVALUES

9.4.1 Debris flows/avalanches

Fabric characteristics for the various process origins are summarised in Table 2.3. There is some variability in the orientation of particles in the Tsatsa-La-Mangaung, Sekhokong, Leqooa Valley and Sehonghong deposits (Figures 5.16, 6.9, 6.27, 7.16, 7.19 and 8.12), which corresponds with the range of debris flow orientations described in section 2.1.2.2.1. Clasts at all sites predominantly dip between 0° and 40° (Figures 5.16, 6.9, 6.27, 7.16, 7.19 and 8.12), which is similar to those described for debris flows (Innes, 1983; Boelhouwers *et al.*, 1998), however there is also a clustering around 60° in trench 2 of the Sekhokong Site 1 deposit (Figure 6.9), which is much higher than typical debris flow dips. The suggestion that imbrication in debris flows may reflect the abundance of matrix and that clast-rich debris flows would show stronger imbrication (Bertran *et al.*, 1997) is observed within the linear slope deposits, where the Tsatsa-La-Mangaung (Figure 5.16), trench 2 of the Sekhokong Site 1 (Figure 6.9) and all trenches of the Sekhokong Site 2 deposit (Figure 6.27) display a more well-developed imbrication than trenches 1 and 3 of the Sekhokong Site 1 deposit and the Leqooa Valley deposits (Figures 7.16 and 7.19), which contain more fine-grained material.

Although the Tsatsa-La-Mangaung deposit has moderate to high elongation values (0.57 to 0.77), it has very low isotropy values (0.06 to 0.14) (Figure 5.17). The Sekhokong Site 1 deposit has very low isotropy values (0.06 to 0.11) and low to moderate elongation values (0.47 to 0.66) (Figure 6.10), which again does not correspond to typical debris flow values (Table 2.3), however isotropy values for the Sekhokong Site 2 deposit range from 0.06 to 0.12, which is slightly lower than typical debris flow isotropy and elongation values are variable, ranging from 0.22 to 0.66 (Figure 6.28). The Leqooa Valley deposits both have very low isotropy values (0.03 to 0.05 for the western deposit and 0.02 to 0.05 for the eastern deposit), whilst elongation values are high (0.74 for the western deposit) and moderate to high (0.49 to 0.69 for the eastern deposit) (Figures 7.17 and 7.20), which again does not conform

to typical debris flow isotropy values (Table 2.3). Isotropy values are low at Sehonghong Sites 2 and 3 (0.03 to 0.05) (Figure 8.13), whilst isotropy within debris flows tends to be moderate. In contrast, elongation values obtained at Sites 2 and 3 fall within the range of debris flows. Isotropy and elongation values at Site 1 fall within the range for debris flows and the sediment falls within the debris flow envelope (Figure 8.13). None of the linear deposits fall within the debris flow envelope of Lawson (1979) and Mills (1984), with the exception of the sediment in trench 1 of the Sekhokong Site 1 deposit. In addition, the sediment within trenches 1 and 2 of the Sekhokong Site 2 deposit fall very close to the debris flow envelopes (Figure 6.29).

9.4.2 Landslides

The parallel or transverse clast orientations observed within landslides (Table 2.3) corresponds to the clast orientations within the Tsatsa-La-Mangaung deposit, which are parallel to movement in the case of the sediment in trenches 1 and 3, and transverse to movement in trench 2 (Figure 5.16). Although clasts in trench 3 of the Sekhokong Site 1 deposit are predominantly oriented transverse to the deposit, clast orientation within trenches 1 and 2 is fairly random (Figure 6.9), which does not concur with clast orientation within landslides (Table 2.3). Clasts within trench 3 of the Sekhokong Site 2 deposit are oriented downslope, which is similar to orientations observed within landslides, however clasts within trenches 1 and 2 are randomly oriented (Figure 6.27) and therefore do not concur with landslide clast orientations. The Leqooa Valley western slope deposit displays similar clast orientations to landslides in the downslope direction of clasts in trenches 1 and 2 and slightly transverse orientations in trench 3 (Figure 7.16). In contrast, clast orientations in trenches 2 and 3 of the eastern deposit are random (Figure 7.19).

9.4.3 Mudflows/mudslides

The Tsatsa-La-Mangaung deposit shows preferred parallel or transverse orientation (Figure 5.16), whilst the Sehonghong Site 3 deposit displays transverse orientations to the hillslope, both of which do not concur with the lack of preferred orientation typically found within mudflow deposits (Table 2.3). In contrast, both the Sekhokong Site 1 and Site 2 deposits display random orientations in trenches 1 and 2 and transverse orientations in trench 3 (Figures 6.9 and 6.27), which is similar to mudflow

fabrics. Clast orientation within the Leqooa Valley western deposit tends to be downslope or transverse to the deposit (Figure 7.16), which does not correspond to the typical random fabrics in mudflows. Alternatively, clast orientation in trenches 2 and 3 of the eastern deposit are random (Figure 7.19) and concur with typical mudflow fabrics. Clasts at Sites 1 and 2 of the Sehonghong deposits are also randomly oriented (Figure 8.12), which is also similar to typical mudflow fabrics.

9.4.4 Solifluction lobes

Clast orientation tends to be in the direction of the slope in trenches 1 and 3 of the Tsatsa-La-Mangaung deposit, in trench 1 of the Leqooa western deposit and at Sehonghong Site 1, which may indicate solifluction processes (Figures 5.16, 7.16 and 8.12). A transverse orientation towards the lower part of the deposit is observed at Sekhokong Site 2 (Figure 6.27) and at the Leqooa western deposit (Figure 7.16) as suggested for solifluction lobes by Benedict (1976). In contrast, clast dips are higher within the linear slope deposits than typical solifluction clast dips, with the exception of the Leqooa western deposit, where clasts tend to dip <30° (Figure 7.16).

Eigenvalues S_1 range between 0.811 and 0.971 for solifluction lobes in Alaska (Nelson, 1985) (Table 2.3), which is significantly higher than those recorded for all deposits (Figures 5.16, 6.9, 6.27, 7.16, 7.19 and 8.12). According to Millar (2005), eigenvalue S_1 results may vary between 0.422 and 0.804 within solifluction deposits, which concurs with the S_1 values for all the deposits in this research which range between 0.526 in trench 1 at Sekhokong Site 1 and 0.813 in trench 1 of the Leqooa western deposit. Gelifluction lobe fabrics have low isotropy values (0.044 to 0.165) and a very wide range of elongation values (0.167 to 0.671) (Benn, 1994a). Isotropy and elongation values for the Tsatsa-La-Mangaung (Figure 5.17), Sekhokong Sites 1 and 2 (Figure 6.10 and 6.28) and Sehonghong Sites 1 and 2 (Figure 8.13) concur with the isotropy and elongation values of gelifluction deposits. In contrast, isotropy values for the Leqooa Valley deposits are slightly lower than typical values for gelifluction lobes, whereas elongation values are slightly higher (Figures 7.17 and 7.20). Clast fabric within trench 2 of the Tsatsa-La-Mangaung deposit (Figure 5.17), all three trenches of the Sekhokong Site 1 deposit (Figure 6.10), trenches 1 and 2 of the Sekhokong Site 2 deposit (Figure 6.28) and Sehonghong Site 1 (Figure 8.13) fall within the gelifluction envelope of Ballantyne (1981).

9.4.5 Rock glaciers

Clasts within trenches 1 and 3 of the Tsatsa-La-Mangaung deposit (Figure 5.16) and in trench 1 of the Leqooa Valley western deposit (Figure 7.16) tend to be oriented downslope, which is similar to clast orientations within a rock glacier described by Harrison and Anderson (2001). The S_1 values from within the rock glacier described by Harrison and Anderson (2001) range from 0.706 to 0.811, which is also similar to those obtained for the Tsatsa-La-Mangaung and Leqooa Valley deposits (Figures 5.16, 7.16 and 7.19). In contrast, the clast orientations within trench 2 of the Tsatsa-La-Mangaung deposit (Figure 5.16), all trenches within the Sekhokong Site 1 deposit (Figure 6.9), trenches 1 and 2 of the Sekhokong Site 2 deposit (Figure 6.27), trenches 2 and 3 of the Leqooa western deposit (Figure 7.16), and all trenches of the eastern deposit (Figure 7.19) concur with clast orientation within a rock glacier described by Vere and Matthews (1985) (Table 2.3).

The S_1 values for a rock glacier described by Giardino and Vitek (1985) range from 0.625 to 0.961, which is similar to those obtained for the Tsatsa-La-Mangaung (Figure 5.18), Sekhokong Site 1 (Figure 6.9) and Leqooa Valley deposits (Figure 7.16 and 7.19). More recent work on eigenvalues and fabric strength within rock glaciers measured S_1 values ranging from 0.4117 to 0.6435 (Nicolas, 1994). These values are similar to those obtained for clasts within trench 2 of the Tsatsa-La-Mangaung and Sekhokong Site 1 (Figure 5.16 and 6.9) deposits and trenches 1 and 2 of the Sekhokong Site 2 deposit (Figure 6.27). Clast dips recorded for the Leqooa western deposit (Figure 7.16) do not concur with typical rock glacier clast dips (Table 2.3), however many clasts within the other linear deposits also have dip values >30°. The sediment in all deposits, with the exception of trenches 1 and 2 of the Sekhokong Site 2 deposit fall within the rock glacier internal fabric envelope based on work by Giardino and Vitek (1985) (Figures 5.18, 6.11, 6.30, 7.18 and 7.22).

9.4.6 Pronival ramparts

The random clast orientations observed within the Sekhokong Site 1 deposit (Figure 6.9), trenches 1 and 2 of the Sekhokong Site 2 deposit (Figure 6.27), trenches 2 and 3 of the Leqooa western deposit (Figure 7.16) and trench 2 of the eastern deposit (Figure 7.19) are similar to orientation characteristics for pronival ramparts (Table 2.3). In contrast, it has also been suggested that an oblique alignment of clasts may

occur in pronival ramparts with downslope dips (Harris, 1986; Shakesby *et al.*, 1999), which also occur in trench 2 of the Tsatsa-La-Mangaung deposit (Figure 5.16), trench 3 of the Sekhokong Site 1 and Site 2 deposits (Figures 6.9 and 6.27) and trench 1 of the Leqooa eastern deposit (Figure 7.19).

9.4.7 Glacial moraine

The weak orientations observed within the Sekhokong Site 1 deposit (Figure 6.9), trenches 1 and 2 of the Sekhokong Site 2 deposit (Figure 6.27) and trenches 2 and 3 of the Leqooa eastern deposit (Figure 7.19) are similar to those of surface sediments in push moraines (Sharp,1984). Had glaciers been present at the linear deposit sites, orientations parallel to glacier flow would concur with clasts within trenches 1 and 3 of the Tsatsa-La-Mangaung deposit (Figure 5.16) and trenches 1 and 2 of the Leqooa western slope deposit (Figure 7.16).

Owen (1994) reports that supraglacial flow tills exhibit weak down-valley clast fabrics, which would conform to the Tsatsa-La-Mangaung (Figure 5.16), Sekhokong Site 2 (Figure 6.27), Leqooa Valley deposits (Figure 7.16 and 7.19) and Sehonghong Sites 1 and 2 deposits (Figure 8.12). Bennett *et al.* (1999) report that the average flow till isotropy value is 0.122 and average elongation value is 0.436 for clasts sampled in Svalbard, whereas Lawson (1979) reported an average isotropy value of 0.231 and elongation value of 0.455 in Alaska. However, such isotropy and elongation values do not concur with those for the deposits cited above (Table 2.3).

The linear and Sehonghong deposits display similar clast orientations to those described for supraglacial melt out till (Table 2.3 and 2.6). Isotropy for supraglacial melt-out till ranges from 0.037 (Lawson, 1979) to 0.209 (Bennett *et al.*, 1999), whilst elongation values range from 0.428 (Bennett *et al.*, 1999) to 0.82 (Lawson, 1979) (Table 2.3). Isotropy and elongation values for the Tsatsa-La-Mangaung, Sekhokong Site 1 and Sehonghong Sites 1 and 2 fall within the range specified for supraglacial melt-out till (Figures 5.17, 6.10 and 8.13), as do clast dip values which are predominantly <40° (Dowdeswell and Sharp, 1986) (Figure 5.16, 6.9 and 8.12).

Isotropy averages 0.114 whilst elongation averages 0.491 in lodgement tills (Bennett *et al.*, 1999) (Table 2.3). Although clasts within the Tsatsa-La-Mangaung (Figure

5.17), Sekhokong Site 1 (Figure 6.10) and Sehonghong Sites 1 and 3 (Figure 8. 13) fall within the undeformed lodgement till envelope, clasts do not show strong preferred orientation in the direction of ice flow. Clast orientation within the Leqooa western deposit (Figure 7.16) is similar to orientations in Subglacial melt-out till and basal ice (Table 2.3 and 2.6). In addition, the sediment in all three trenches falls within the basal ice envelope (Figure 7.17). In contrast, clasts within the Leqooa eastern deposit in trenches 1 and 3 (Figure 7.20) also fall within the basal ice envelope; however only clasts in trench 1 display strong preferred orientation (Figure 7.19). Clasts at Sehonghong Site 3 also display strong orientations (Figure 8.12) and fall within the basal ice envelope (Figure 8.13). Steeply dipping clasts are common in basal ice deposits (Dowdeswell and Sharp, 1986); however clasts at the sites described above do not dip particularly steeply. Isotropy ranges from 0.038 (Ham and Mickelson, 1994) to 0.097 (Bennett *et al.,* 1999), whilst elongation values range from 0.597 (Bennett *et al.,* 1999) to 0.903 (Lawson, 1979) for basal ice, which is similar to values obtained for the Leqooa western deposit (Figure 7.17) and Sehonghong Site 3 (Figure 8.13).

9.4.8 Fluvial deposits

Clasts sampled from the south-facing deposits display random orientations which are not characteristics of fluvial deposits; however those sampled from the north-facing deposit are oriented parallel to the flow of the Sehonghong River (Figure 8.12). In contrast, clast imbrication at Sehonghong Site 3 is downstream, suggesting that fabric at this site is not characteristic of fluvial deposits.

9.4.9 Summary

The difficulty in using fabric orientation in the interpretation of a process origin, as described in section 2.2.2, is represented in this study. Clast orientations within all the Drakensberg deposits display characteristics from numerous processes and cannot be used alone to infer an origin.

9.5 PARTICLE SHAPE

9.5.1 Debris flows/avalanches

A summary of particle shape for various deposits is presented in Table 2.4. The predominantly angular clasts sampled in the linear deposits and at Sehonghong Site 1

and 2 (Figures 5.22, 6.15, 6.33, 7.25, 7.29 and 8.18) are similar to the angular nature of clasts within debris flows described in section 2.1.2.2.1. The shape of debris flow clasts have been recorded as platy (Johnson and Rodine, 1984), whereas clast shape within all deposits is predominantly bladed (Tables 5.7, 6.4, 6.15, 7.7, 7.12 and 8.4). In contrast, although bladed shapes form the modal class in trench 3 of the Sekhokong Site 2 deposit, the Leqooa deposit and Sehonghong Site 3, there is a considerable proportion of platy clasts present at these sites. Clasts within all deposits with the exception of Sehonghong Site 3 are predominantly angular, which may also reflect debris avalanche processes.

9.5.2 Landslides and mudflows/mudslides

The angular nature of clasts within landslides as described in section 2.1.2.2.3 is similar to clast roundness at all sites, however the proportion of clasts falling in the very angular and angular categories for landslides described by Hewitt (1999) is significantly higher than clasts falling within this category of angularity at all sites (Tables 5.9, 6.6, 6.17, 7.9, 7.14 and 8.6). Clast shape within mudflows described in section 2.1.2.2.4 is also angular, which is similar to clast shape within the linear slope deposits and Sehonghong Sites 1 and 2 (Figures 5.22, 6.15, 6.33, 7.25, 7.29 and 8.18).

9.5.3 Solifluction lobes

Clasts within all linear deposits predominantly fall in the very angular to subangular classes of roundness (Tables 5.9, 6.6, 6.17, 7.9 and 7.14), which is similar to the angular nature of clasts within solifluction deposits. However, the predominant clast shape within all linear deposits is bladed (Tables 5.7, 6.4, 6.15, 7.7 and 7.12), which does not concur with the platy nature of clasts within solifluction deposits (Table 2.4). Clasts at Sites 1 and 2 of the Sehonghong deposit are also predominantly angular (Table 8.6), however clasts at Site 3 are only angular in the upper units in the coarse layer, whilst below this layer, clasts are more rounded, which does not concur with that of typical solifluction deposits. Bladed shapes form the modal class at all three sites (Table 8.4), which again does not conform to the platy nature of clasts within solifluction landforms.

9.5.4 Rock glaciers

The predominantly platy nature of clasts within rock glaciers does not concur with the predominantly bladed clasts within the linear slope deposits (Tables 5.7, 6.4, 6.15, 7.7 and 7.12). The C_{40} ratio for the rock glacier in the Nantlle Valley, North Wales, is between 40 and 66 (Harrison and Anderson, 2001) (Table 2.4), which is similar to the C_{40} values for the Tsatsa-La-Mangaung deposit (Figure 5.19), trenches 1 and 2 of the Sekhokong Site 1 deposit (Figure 6.12), the Sekhokong Site 2 deposit (Figure 6.30), trench 2 of the Leqooa western deposit (Figure 7.22) and trench 1 of the Leqooa eastern deposit (Figure 7.26). However, the C_{40} ratio of as low as 20 has been recorded for a rock glacier in Norway (Vere and Matthews, 1985). Such a low value may however reflect subglacial processes, given that the rock glacier formed from a lateral moraine (Vere and Matthews, 1985). This lower C_{40} value is similar to that obtained within trench 3 of the Sekhokong Site 2 deposit (Figure 6.30). The angular nature of clasts within rock glaciers (Table 2.4) fits the description of clasts located at all the linear deposit sites (Tables 5.9, 6.6, 6.17, 7.9 and 7.14).

9.5.5 Pronival ramparts

Clast shape within the Tsatsa-La-Mangaung deposit (Figure 5.22), trench 1 of the Sekhokong Site 1 deposit (Figure 6.15), trench 1 of the Leqooa western deposit (Figure 7.25) and all trenches within the eastern deposit (Figure 7.29) have an angular modal class and small amount of subrounded clasts, which is similar to particle shape of pronival ramparts (Table 2.4). In contrast, clasts within trench 3 of the Sekhokong Site 1, Site 2 and Leqooa western deposit amount to 28%, 30% and 24% subrounded clasts respectively (Tables 6.6, 6.17 and 7.9), which is a significantly higher percentage than those found within pronival ramparts. Clast shape within pronival ramparts has been described as slabby, with a C_{40} index of ~38 to ~69 (Ballantyne and Kirkbride, 1986) (Table 2.4). The C_{40} values obtained for the clasts within the Tsatsa-La-Mangaung deposit (Figure 5.19), trenches 1 and 2 of the Sekhokong Site 1 deposit (Figure 6.12), the Sekhokong Site 2 deposit (Figure 6.30) and trench 1 of the Leqooa eastern deposit (Figure 7.26) also fall within the range suggested by Ballantyne and Kirkbride (1986).

9.5.6 Glacial moraines

The predominantly angular and bladed nature of the clasts sampled from the linear deposits (Figures 5.22, 6.15, 6.33, 7.25 and 7.29; Tables 5.7, 6.4, 6.15, 7.7 and 7.12) is similar to the nature of supraglacial debris (Table 2.4). This is also true of clasts sampled within the upper units of the Sehonghong south-facing and north-facing deposits (Figure 8.18 and Table 8.4). The C_{40} indices for supraglacial till (Benn and Ballantyne, 1993) are above 60, whilst samples collected for till from Slettmarkbreen in Norway have a C_{40} index below 16 (Benn and Ballantyne, 1993) (Table 2.4). C_{40} values obtained for the Tsatsa-La-Mangaung deposit (between 46 and 54) (Figure 5.19) fall very close to values obtained for supraglacial till, whilst values for the remaining linear deposits fall between the supraglacial till and till values (Figures 6.12, 6.30, 7.22 and 7.26), possibly reflecting both active and passive processes. Clasts at Sehonghong Sites 1 and 2 also fall between values for supraglacial till and till, whilst clasts sampled at Site 3 fall within the range for till (Figure 8.15). All the linear deposits display a slight downslope tendency of increased clast roundness (Figures 5.22, 6.15, 6.33, 7.25 and 7.29).

9.5.7 Fluvial deposits

The percentage of spherical shapes found in fluvial deposits and described in section 2.1.6 are considerably higher than at all the Sehonghong sites (Table 8.4), where they make up between 0% and 4% of the total percentage of clasts. However, Figure 8.15 indicates that the majority of clasts fall within the compact categories of bladed, elongate and platy clasts and therefore indicate a certain degree of rounding. Prolate (rod-shaped/elongate) clasts are characteristic of river gravels (Howard, 1992), which is also indicated by the relatively high proportions of elongate clasts (32% and 36%) at Sites 1 and 3 respectively (Table 8.4). Subrounded clasts comprise 28% of the total sample at Site 3 (Table 8.6), which is also similar to fluvial deposits. In contrast, subrounded clasts make up between 0% and 8% of the total sample at Sites 1 and 2 (Table 8.4). Tributary sediments have been recorded as having RA indices (percentage of clasts in the very angular and angular categories) of 30% to 50% (McEwen and Matthews, 1998), which is similar to RA values at Sites 1 and 3 (46% and 18% respectively). Although clasts at Sites 1 and 3 display fairly rounded shapes, these may simply indicate greater amounts of weathering as discussed previously regarding the roundness of basalt clasts.

The maximum projection sphericity best reflects the hydraulic behaviour of particles and there is a close relationship between maximum projection sphericity and the velocity at which a particle of particular volume will settle in a fluid or roll along a bed (Gale and Hoare, 1991). Increase in sphericity of clasts in rivers occurs at a slow, constant rate (Bridge, 2003). Mean maximum projection sphericity at Sites 1 and 3 falls between 0.7 and 0.8, whilst mean maximum projection sphericity at Site 2 is 0.6 (Figure 8.17), indicating that clasts at Sites 1 and 3 are more spherical and would therefore preferentially roll along a bed.

9.5.8 Summary

As in the case of clast orientation, clast shape cannot be an indicator of process origin based on the wide range of processes which display similar clast shape characteristics to the Drakensberg deposits. Lithology does not appear to vary considerably at the sites; therefore the difference in clast shape between sites is believed to reflect different transport processes and will be discussed further in later sections.

9.6 MICROMORPHOLOGY

9.6.1 Debris flows, landslides and mudslides

Trench 3 of the Sekhokong Site 1 deposit (Figure 6.18) and trench 2 of the Leqooa western deposit (Figure 7.31) contain rotational features, which also occur within debris flow deposits as described in section 2.1.2.2.1. There is also evidence for preferential orientation of clasts along a shear within trench 3 of the Sekhokong Site 1 deposit (Figure 6.18), which again may occur within debris flow deposits. The Sekhokong Site 2, Leqooa eastern and Sehonghong Site 1 deposits display a very coarse-grained structure (Figure 6.34), which may be due to the movement of fluid, which has the ability to wash away fines, leaving zones of sand-sized or larger particles (Van der Meer, 1993). In contrast, the Tsatsa-La-Mangaung, Leqooa eastern and Sehonghong Site 2 deposits do not show any micromorphological features associated with debris flows (Tables 5.11, 6.20, 7.18 and 8.9), and the silt coatings which do occur within the deposits are not disturbed and scattered within the matrix, as can occur within debris flows (Van Vliet-Lanoë, 1985a) (Figures 5.28 and 8.19). However, it must be taken into account that most work has been undertaken on fine-grained debris flows, whilst the deposits sampled in this research are coarse-grained. It remains difficult to sample coarse-grained deposits without destroying the

microstructure, and in such cases macroscopic microstructures are the only ones which can be used to assess the origin of the deposits (Bertran, Pers Comm., 2006). Although recent work on the micromorphology of coarse-grained debris flows has been undertaken, such information remains unpublished (Menzies, Pers Comm., 2006). The deposits do not contain slip planes or other microfeatures associated with slides (Tables 5.11, 6.9, 6.20, 7.17, 7.18 and 8.9).

9.6.2 Solifluction lobes

Platy structures are observed within all micromorphological sections collected within the Tsatsa-La-Mangaung (Table 5.11) and Sekhokong Site 1 (Table 6.9) deposits and are typical of solifluction deposits as described in section 2.2.1.6. Further features typical of solifluction occur at Sehonghong Site 2 in the form of clay coatings (Figure 8.20 and 8.21). However these clay coatings are only at 100 cm depth and therefore probably represent postdepositional illuviation of clay. Silt cappings on the surface of clasts occur within all deposits to a certain extent (Tables 5.11, 6.9, 6.20, 7.17, 7.18 and 8.9). In addition, cappings become more frequent downslope within the Tsatsa-La-Mangaung deposit, which is once again indicative of solifluction deposits. Encircling silt cappings surround the more rounded particles at Sehonghong Site 1 (Figure 8.19), which indicates that gelifluction processes may have occurred at this site (Van Vliet-Lanoë, 1985b, 1987).

9.6.3 Glacial moraines

Discrete shears, rotation structures and pressure shadows are present in samples collected from the Sekhokong Site 1 deposit and the Leqooa western deposit (Figures 6.18 and 7.31), which are also found within glacial deposits (section 2.1.5). The thin sections obtained from within the Tsatsa-La-Mangaung, Sekhokong Site 2 and Leqooa eastern deposit indicate that the sediment is not overconsolidated, which may reflect supraglacial properties.

9.6.4 Fluvial deposits

Although clay coatings lining voids at Sehonghong Site 2 at 100 cm depth are present (Table 8.9, Figures 8.20 and 8.21), these are believed to be a function of present-day pedogenic processes. Unfortunately, fluvial micromorphological research appears to

be restricted to palaeosols within floodplains, and thus no comparisons can be made between samples collected in this research with those from fluvial environments.

9.6.5 Summary

The majority of the thin sections collected from the deposits display silt cappings, which is typical of solifluction processes. The Sekhokong Site 2, Leqooa eastern and Sehonghong Site 2 deposits display a coarse-grained structure which could be associated with debris flow processes. In contrast, the Sekhokong Site 1 and Leqooa western deposits contain several rotational structures which may be associated with debris flow processes or glacial processes. A major problem in ascribing a process origin based on micromorphological analysis arises from the apparent similarities between sediment flow deposits, solifluction deposits and subglacially deformed deposits, such as for example in the presence of discrete shears, rotation structures and cappings. It is therefore important that microstructural analyses be undertaken in conjunction with other sedimentological analyses (Van der Meer, 1993; Lachniet *et al.*, 2001).

9.7 EROSION REMNANT

A final process origin which has not previously been considered is that the linear deposits may be debris mantles which have been fluvially incised to produce the ridge morphologies. This process origin can be discarded based on a comparison between the particle size characteristics of the Tsatsa-La-Mangaung deposit and the pits sampled on the adjacent slope (Figures 5.12 to 5.14 and 5.15). There is a clear difference in texture between the pits and the deposit, with the pits containing a much greater amount of sand and silt than the material within the deposit. Pits 1 and 5 are described as containing more organic matter than the other pits (Figures 5.7 and 5.11). This corresponds with the statement by Schmitz and Rooyani (1987) that organic soils have been identified at elevations above 3000 m a.s.l., as pits 1 and 5 occur between 3090 m a.s.l. and 3100 m a.s.l. (Figure 5.2). These organic soils form under a dense vegetation cover where temperatures are below 8°C and where drainage is impeded (Schmitz and Rooyani, 1987). Pits 3 and 4 are typical of the shallow soils found on the mountain slopes of Lesotho as a result of erosion and low temperatures (Schmitz and Rooyani, 1987), whereas pit 2 strongly resembles the sediment found within the Tsatsa-La-Mangaung deposit (Figures 5.4 to 5.6 and 5.8). This may be explained by

the presence of a small cirque glacier covering the area where pit 2 is situated. This cirque glacier would have been moving over this area in order to supply material to the deposit and some may have been deposited on retreat of the glacier.

It has been suggested by Sumner (2004c), that the Tsatsa-La-Mangaung deposit is a relict openwork block accumulation and that deposits ascribed to glacial processes by Grab (1996a) appear to be incised colluvial mantles. The fluvially incised debris mantle is discarded as a possible process origin on the basis that it would be expected that the material within the deposit would be similar to that on adjacent slopes if it had simply been fluvially incised *in situ*. In addition, the colluvial material observed at Sehonghong is significantly different to the sediment within the linear slope deposits. Although the deposits contain a significant amount of blocks within them, these are not regarded as openwork deposits, given that the blocks are matrix supported.

9.8 POSSIBLE PROCESS ORIGIN FOR THE TSATSA-LA-MANGAUNG DEPOSIT

It is evident from the discussion in section 2.2, that there can be great difficulty in identifying and distinguishing between various process mechanisms that occur within the periglacial and glacial zone. Great care must therefore be exercised in identifying a possible process origin for the Tsatsa-La-Mangaung deposit, and all possible hypotheses should be explored.

In order to help identify a possible process origin for the Tsatsa-La-Mangaung deposit, all morphological, sedimentological and micromorphological characteristics have been summarised to determine whether they are typical attributes of the possible process mechanisms suggested for the formation of the deposit (Table 9.1). Although lithology influences the various sedimentary characteristics investigated in this research, morphological and micromorphological characteristics are also taken into account when determining a process origin. The suggestion that the Tsatsa-La-Mangaung deposit may be a debris flow or debris avalanche can be rejected on the basis that it does not display the necessary attributes in terms of morphology, fabric, eigenvalue, particle shape or micromorphology. Similarly, a mudflow/mudslide and landslide origin can be rejected as the Tsatsa-La-Mangaung deposit does not host the attributes regarded as characteristic of these features. The deposit hosts only the

Table 9.1 Suitability of geomorphic processes and landform types for the various Tsatsa-La-Mangaung deposit attributes.

	Morph-ology	Particle Size	Sorting	Fabric	Eigen-value	Clast shape	Round-ness	Micro-morphology	
Geomorphic Process									*Geomorphic Landform*
Debris flow	✗	✓	✓	✓	✗	✗	✓	✗	Debris deposit
Mass movement	✓	✗	?	✓	?	?	✓	✗	Landslide
Mudflow/slide	✗	✗	?	✗	?	?	✓	✗	Mudflow/mudslide
Solifluction	✗	✗	✗	✓	✓	✗	✓	✓	Solifluction lobe
Periglacial/permafrost	✗	✗	✓	✓	✓	✗	✓	?	Rock Glacier
Nival	✓	✓	✓	✗	?	✓	✓	?	Pronival rampart
Glacial	✓	✓	✓	✓	✓	✓	✓	?	Moraine

sorting, fabric and eigenvalue attributes of a rock glacier and the fabric, eigenvalue and roundness characteristics of a solifluction landform, and is therefore not considered to be either of these features. However, the Tsatsa-La-Mangaung deposit hosts most of the morphological and sedimentological characteristics typical to pronival ramparts, with the exception of the fabric results.

The fabric patterns are dismissed in this case, based on the discussion in section 2.2.2, where it is suggested that fabric may not be truly representative of a facies under investigation (Benn and Ringrose, 2001), as it can be affected by clast properties (Glenn *et al.,* 1957; Boulton, 1970; Drake, 1974; Kjær and Krüger, 1998; Carr and Rose, 2003; Millar, 2005). Although the Tsatsa-La-Mangaung deposit shows several characteristics typical to pronival ramparts, such ramparts usually form parallel to the backwall and at the foot of talus slopes overlooking rockwalls (Lewis, 1966; Linder and Marks, 1984). This is not true for many end and lateral moraines (Ballantyne and Kirkbride, 1986), neither is it true for the Tsatsa-La-Mangaung deposit. It is thus suggested that the Tsatsa-La-Mangaung deposit is unlikely to be a fossil pronival rampart.

It would appear that the Tsatsa-La-Mangaung deposit hosts all the morphological and sedimentological characteristics typical to glacial moraines. Although the Tsatsa-La-Mangaung deposit does not possess the micromorphological attributes of glacial deposits, there is no documented literature on the micromorphological characteristics of supraglacial deposits, and thus no comparison can be made in this regard. The Tsatsa-La-Mangaung deposit displays the micromorphological attributes and fabric and roundness characteristics of a solifluction lobe; however it does not demonstrate any other morphological or sedimentological features typical to solifluction lobes. It is suggested that post depositional changes may include the translocation of clay and silt and this process is usually visible as a coating (Van der Meer, 1987). It is thus proposed that post-depositional solifluction processes have taken place within the Tsatsa-La-Mangaung deposit, as is reflected by the typical solifluction fabrics and micromorphology.

The sedimentary characteristics of the Tsatsa-La-Mangaung deposit resemble those of supraglacial melt-out till and flow till. The clasts within the Tsatsa-La-Mangaung

deposit are predominantly angular (Figure 5.22 and Table 5.9) and not striated or faceted. The grain-size distribution is coarse and unimodal (Figures 5.12 to 5.14, 9.5 and Table 5.1), clast fabric is variable (Figure 5.17), and particles are poorly consolidated and the structure is massive (Figures 5.4 to 5.6). These sedimentary results imply that the Tsatsa-La-Mangaung deposit is a moraine, which can be further quantified by comparing C_{40} index and roundness characteristics of the Tsatsa-La-Mangaung deposit to envelopes produced by Benn and Ballantyne (1994) on a covariance plot (Figure 9.6). The clasts sampled in trenches 1 and 3 fall within the moraine envelope, whilst the clasts sampled in trench 2 fall just outside this envelope. The moraine envelope represents clasts sampled from lateral and end moraines in Jotunheimen, Norway (Benn and Ballantyne, 1994). Results from the lateral moraines are similar to those for the Tsatsa-La-Mangaung deposit. High C_{40} values (around 44) and high RA values (percentage of clast falling within the very angular and angular categories) (between 44 and 62) are typical for lateral moraines, whilst the Tsatsa-La-Mangaung deposit has C_{40} values ranging between 46 and 54 (Figure 5.19) and RA values between 58 and 62 (Figure 5.22 and Table 5.9). These values are significantly higher than those recorded by Benn and Ballantyne (1994) for the end moraines, which may be as a result of clast modification during transport, or a progressive increase in the proportion of actively transported clasts towards the glacier snout producing lower C_{40} and RA values (Benn and Ballantyne, 1994).

The Tsatsa-La-Mangaung deposit displays similar morphological and sedimentological characteristics to lateral moraines. It displays evidence of passive transport in terms of its particle size analysis, which is predominantly coarse-grained; however there is also evidence of active transport through the presence of subangular clasts. It is therefore suggested, based on the morphological and sedimentological characteristics, that the Tsatsa-La-Mangaung deposit is a moraine.

9.9 POSSIBLE PROCESS ORIGIN FOR THE SEKHOKONG SITE 1 DEPOSIT

In order to help identify a possible process origin for the Sekhokong Site 1 deposit, all morphological, sedimentological and micromorphological characteristics have been summarised, in terms of whether they are typical attributes of the possible process

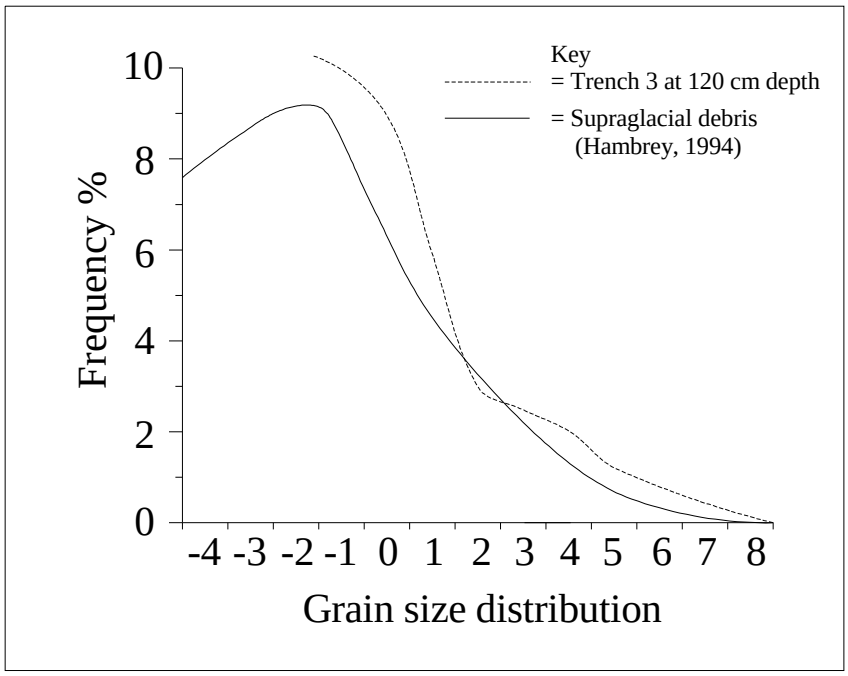


Figure 9.5 Typical grain size distributions for supraglacial debris compared to an example at Tsatsa-La-Mangaung (after Hambrey, 1994).

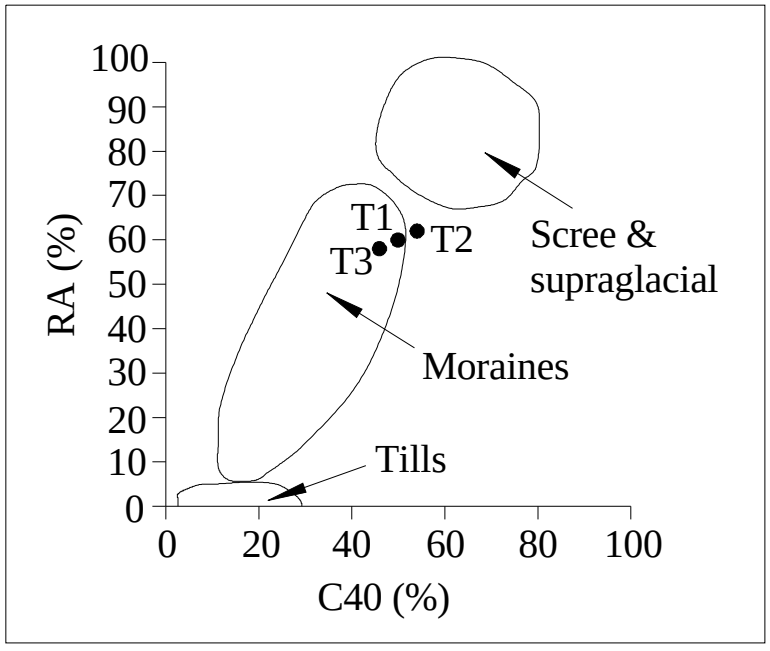


Figure 9.6 Plotting the Tsatsa-La-Mangaung sediments against the co-variance plot of C_{40} and RA indexes and sediment envelopes (after Benn and Ballantyne, 1994) (T1 = trench 1, T2 = trench 2, T3 = trench 3).

mechanisms suggested for the formation of the deposit (Table 9.2). The theory that the Sekhokong Site 1 deposit may be a debris flow or avalanche can be rejected on the basis that it does not possess the necessary attributes in terms of morphology, eigenvalue or particle shape. Similarly, a mudflow/mudslide and landslide origin can be rejected as the Sekhokong Site 1 deposit displays only the fabric and roundness characteristics of a mudflow or mudslide and only the fabric characteristics of a landslide. The deposit possesses only the sorting, eigenvalue and roundness attributes of a rock glacier and the eigenvalues and clast shape of a solifluction lobe, and is therefore unlikely to represent any such fossil feature.

It would appear that the Sekhokong Site 1 deposit hosts all the morphological and sedimentological characteristics typical to a moraine. The Sekhokong Site 1 deposit also hosts all the morphological and sedimentological characteristics typical to a pronival rampart, with the exception of the roundness results. Although the Sekhokong Site 1 deposit shows several characteristics typical to pronival ramparts, the roundness characteristics of the deposit suggest some form of active transport. Although subnival transport in the form of debris flow sediment emerging from the snow bed is accepted to account for the presence of subangular clasts on pronival ramparts (Shakesby *et al.*, 1995), the clasts within the Sekhokong Site 1 deposit are more rounded than that for typical pronival ramparts. Furthermore, pronival ramparts usually form parallel to the backwall and at the foot of talus slopes overlooking rockwalls (Lewis, 1966; Linder and Marks, 1984), whilst the Sekhokong site 1 deposit is positioned downslope. It is thus proposed that the deposit is more likely to represent a glacial moraine than a fossil pronival rampart.

The various transport paths through a glacier have been described in section 2.1.5.2. Based on the sedimentary characteristics of the Sekhokong Site 1 deposit, it is suggested that both active and passive transport processes have taken place in the formation of the moraine. This deposit contains more fines than the Tsatsa-La-Mangaung deposit despite their similar lithology. The sediment is bimodal and clasts are more rounded. The bimodal particle size characteristics of trenches 1 and 2 are typical of subglacially transported sediment. This is evident in Figure 9.7, which indicates the grain-size distribution of subglacially transported debris compared to an example in trench 1 of the Sekhokong Site 1 deposit. Both grain-size distributions

Table 9.2 Suitability of geomorphic processes and landform types for the various Sekhokong Site 1 deposit attributes.

	Morph-ology	Particle Size	Sorting	Fabric	Eigen-value	Clast shape	Round-ness	Micro-morphology	
Geomorphic Process									*Geomorphic Landform*
Debris flow	✗	✓	✓	✓	✗	✗	✓	✓	Debris deposit
Mass movement	✗	✗	?	✓	?	?	✗	✗	Landslide
Mudflow/slide	✗	✗	?	✓	?	?	✓	✗	Mudflow/mudslide
Solifluction	✗	✗	?	✗	✓	✓	✗	✓	Solifluction lobe
Periglacial/permafrost	✗	✗	✓	✗	✗	✗	✓	?	Rock Glacier
Nival	✓	✓	✓	✓	?	?	✗	?	Pronival rampart
Glacial	✓	✓	✓	✓	✓	✓	✓	✓	Moraine

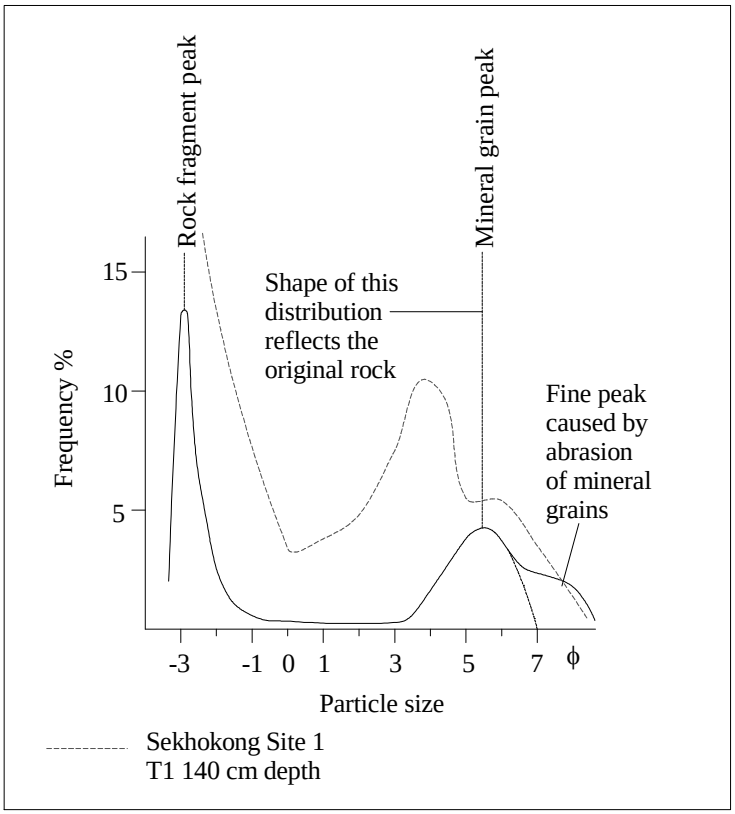


Figure 9.7 The grain-size distribution of subglacially transported debris, composed of three distinct populations: (1) lithic or rock fragments, (2) mineral grains produced by crushing of the rock fragments, (3) fines produced by the abrasion of individual mineral grains (after Bennett and Glasser, 1996), compared to an example at Sekhokong Site 1.

have a primary peak of coarse material and a second mineral grain peak. It is therefore suggested that the sediment at this site underwent a greater amount of active transport than the sediment at the Tsatsa-La-Mangaung deposit. In terms of assigning a type of till to the sediment, the Sekhokong Site 1 deposit contains attributes characteristic of subglacial melt-out till and supraglacial melt-out till (Table 2.6).

The sediment in trench 1 contains angular clasts (Figure 6.15 and Table 6.6) typical of supraglacial melt-out till, however particle size distribution is bimodal (Figure 6.6 and Table 6.1), which is characteristic of subglacial melt-out till. Particle fabric is poorly developed (Figure 6.9) as in the case of supraglacial melt-out till, however the sediment is well packed (Figure 6.3), which is typical of subglacial melt-out till. The sediment is massive (Figure 6.3), which is characteristic of both types of till. In contrast, clasts in trench 2 are more rounded (Figure 6.15 and Table 6.6), which is

characteristic of subglacial melt-out till, however the particle distribution is unimodal (Figure 6.7 and Table 6.1), which it typical of supraglacial melt-out till. The sediment in trench 3 contains all the attributes of subglacial melt-out till, and where clasts are more rounded (Figure 6.15 and Table 6.6), the particle distribution is bimodal (Figure 6.8 and Table 6.1), with the exception of the sample collected at 80 cm depth in trench 3, fabric is oriented transverse to the deposit (Figure 6.9) and the sediment is well packed and massive (Figure 6.5).

The sedimentary results indicate that the Sekhokong Site 1 deposit is likely to be a moraine composed of material which has undergone active and passive transport, which includes lateral moraines. This can be further quantified by comparing C_{40} index and roundness characteristics of the Sekhokong Site 1 deposit to envelopes produced by Benn and Ballantyne (1994) on a covariance plot (Figure 9.8). The clasts sampled in the three trenches fall within the moraine envelope from lateral and end moraines in Jotunheimen, Norway (Benn and Ballantyne, 1994). Results from the lateral moraines are similar to those for the Sekhokong Site 1 deposit, which has C_{40} values ranging between 22 and 48 (Figure 6.12) and RA values between 36 and 60 (Figure 6.15 and Table 6.6).

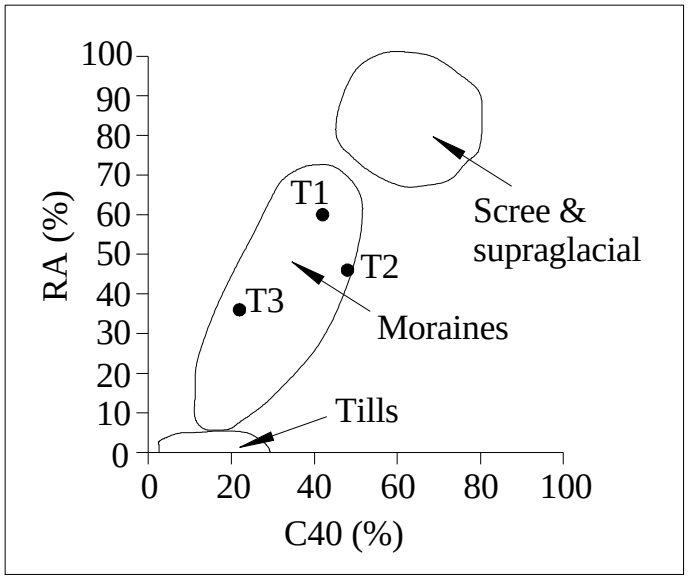


Figure 9.8 Co-variance plot of C_{40} and RA indexes and sediment envelopes (after Benn and Ballantyne, 1994) in comparison to sediment from the three trenches at the Sekhokong Site 1 deposit.

Lateral moraine debris may be predominantly passively transported rockfall material which suffers little modification, retaining its primary parent material characteristics which tend to be coarse, angular clasts with little matrix, however actively transported debris may also take place and result in matrix-rich debris in some dump moraines (Small, 1983; Benn and Evans, 1998; Huddart, 2004). The Sekhokong Site 1 deposit displays similar morphological and sedimentological characteristics to lateral moraines. It displays evidence of passive transport in terms of its particle size distribution, which is predominantly coarse-grained in trench 2, however there is also evidence of active transport through the presence finer material in trenches 1 and 3 and the presence of subangular clasts. It is therefore suggested, based on the morphological and sedimentological characteristics, that the Sekhokong Site 1 deposit is a moraine.

9.10 POSSIBLE PROCESS ORIGIN FOR THE SEKHOKONG SITE 2 DEPOSIT

The morphological, sedimentological and micromorphological characteristics of the Sekhokong Site 2 deposit have been summarised in terms of whether they are typical attributes of the possible process mechanisms suggested for the formation of the deposit (Table 9.3). The suggestion that the deposit may be a mudslide can be rejected given inappropriate morphological and particle size characteristics. The deposit contains none of the typical characteristics associated with landslides. Similarly, a pronival origin may be rejected due to the lack of morphological and clast roundness characteristics. The Sekhokong Site 2 deposit has no morphological, particle size, fabric or eigenvalue characteristics of a rock glacier and only has the eigenvalue characteristics of a solifluction lobe. Similarly, the deposit displays no sorting or eigenvalue characteristics typical of a moraine. Although it contains similar morphological properties, the topographic setting of the area is not conducive to the formation of a moraine, given that the position of the deposit is central to its surrounding topography. Had a glacier occupied this area it is unlikely that it would have produced a moraine in such a central location. In contrast, the deposit displays most of the attributes of a debris flow, which suggests that the process origin of the Sekhokong Site 2 deposit is most likely associated with those producing debris flows.

Table 9.3 Suitability of geomorphic processes and landform types for the various Sekhokong Site 2 deposit attributes.

	Morph-ology	Particle Size	Sorting	Fabric	Eigen-value	Clast shape	Round-ness	Micro-morphology	
Geomorphic Process									*Geomorphic Landform*
Debris flow	✔	✔	✔	✔	✔	✖	✔	✔	Debris deposit
Mass movement	✖	✖	?	✖	?	?	✖	✖	Landslide
Mudflow/slide	✖	✖	✔	✔	?	?	✔	✖	Mudflow/mudslide
Solifluction	✖	✖	?	✖	✔	✖	✖	✔	Solifluction lobe
Periglacial/permafrost	✖	✖	✔	✖	✖	✔	✔	?	Rock Glacier
Nival	✖	✔	✔	✔	?	✔	✖	?	Pronival rampart
Glacial	✔	✔	✖	✔	✖	✔	✔	✖	Moraine

The theory of a debris flow forming the Sekhokong Site 2 deposit is somewhat problematic in terms of the lack of available material for mobilization discussed in section 9.2.1, and the discrepancy in terms of typical clast shape within debris flows. In order to account for sufficient material to form the debris flow, it is suggested that a small glacier occupied the area directly above the deposit, which also explains the unconventional particle shape within the debris flow, which is typical to that of a glacial moraine. Sediment may accumulate in the glacier foreland, which is then subjected to ice pushing and compression by an advancing or re-advancing glacier to create push moraines (Benn and Evans, 1998; Van der Wateren, 2002). These are small moraine ridges, usually less than 10 m in height, produced by minor glacial advances (Benn and Evans, 1998). Material is bulldozed to form a ridge by sweeping up the supraglacial debris which has been dumped from the ice margin, or deformation of proglacial sediment (Bennett and Glasser, 1996).

As discussed previously, the Sekhokong Site 2 deposit displays the particle size and clast shape characteristics typical of both active and passive glacial transport, therefore it is suggested that sediment accumulated at the lower margin of the glacier, either as a push moraine, or simply an accumulation of material resulting from glacier advances during the winter months, which was later released on deglaciation, creating a debris flow. It is suggested that the debris flow was initiated as a result of melt-water, which would indicate that the sediment within the Sekhokong Site 2 deposit may reflect typical attributes of flow till. If water released during ice-melt is not easily displaced, then glacial sediment may become liquefied and flow (Whiteman, 2002). The ratio of sediment stress to shear stress determines whether or not flow occurs. The stress imposed on the debris is increased by the thickening of the debris pile, either by addition of further melt-out at the base, or by the superimposition of debris flows from higher up the slope (Whiteman, 2002). Glacier ice typically steepens towards its margins, which provides locations for a wide variety of mass movement processes to occur, however flow is generally the dominant process as a result of excessive quantities of water (Whiteman, 2002).

Based on the morphological and sedimentological analysis of the Sekhokong Site 2 deposit, and the sedimentological characteristics of both debris flows and glacial till, it is suggested that a small glacier occupied the area directly above the deposit,

pushing and depositing material in its foreland and initiating a debris flow and subsequent deposition on deglaciation.

9.11 POSSIBLE PROCESS ORIGIN FOR THE LEQOOA VALLEY DEPOSITS

All morphological, sedimentological and micromorphological characteristics have been summarised for the Leqooa Valley deposits, in terms of whether they are typical attributes of the possible process mechanisms suggested for the formation of the deposit (Tables 9.4 and 9.5). The western slope deposit does not possess the necessary morphology, eigenvalue or clast shape characteristics of a debris flow or debris avalanche, and is therefore not considered to have originated from such processes. Similarly, a mudflow/mudslide and landslide origin can be rejected as the western slope deposit displays none of the characteristics of a mudflow or mudslide, and only the morphology and fabric characteristics of a landslide. Although the western slope deposit possesses the morphology, particle size and sorting of a pronival rampart and the sorting, fabric, eigenvalue and roundness attributes of a rock glacier, none of the other attributes typical to these features are recorded. In addition, the western slope deposit does not display the morphology, eigenvalue and shape characteristics of solifluction lobes.

The eastern slope deposit possesses the particle size distribution, sorting, fabric and clast roundness characteristics typical of debris flows, however only the fabric and roundness characteristics typical of mudflows and mudslides, and none of the necessary characteristics of landslides. In addition, the eastern slope deposit does not possess the necessary particle size, or clast shape characteristics of a pronival rampart, nor the morphology, particle size or clast shape of a rock glacier. Similarly, a solifluction origin for the eastern slope deposit may be rejected as it does not possess the morphology, eigenvalue and clast shape characteristics typical of these features.

It would appear that the Leqooa Valley deposits host all the morphological and sedimentological characteristics typical to a moraine. Although the eastern deposit does not possess the micromorphological characteristics of glacial processes, these cannot be ruled out as only two thin sections were collected in the deposit and these may not be representative of the deposit as a whole. It is suggested that both active

Table 9.4 Suitability of geomorphic processes and landform types for the various Leqooa western deposit attributes.

	Morph-ology	Particle Size	Sorting	Fabric	Eigen-value	Clast shape	Round-ness	Micro-morphology	
Geomorphic Process									*Geomorphic Landform*
Debris flow	✗	✓	✓	✓	✗	✗	✓	✓	Debris deposit
Mass movement	✓	✗	?	✓	?	?	✗	✗	Landslide
Mudflow/slide	✗	✗	?	✗	?	?	✗	✗	Mudflow/mudslide
Solifluction	✗	✗	✓	✓	✗	✗	✓	✓	Solifluction lobe
Periglacial/permafrost	✗	✗	✓	✓	✓	✗	✓	?	Rock Glacier
Nival	✓	✓	✓	✗	?	✗	✗	?	Pronival rampart
Glacial	✓	✓	✓	✓	✓	✓	✓	✓	Moraine

Table 9.5 Suitability of geomorphic processes and landform types for the various Leqooa eastern deposit attributes.

	Morph-ology	Particle Size	Sorting	Fabric	Eigen-value	Clast shape	Round-ness	Micro-morphology	
Geomorphic Process									*Geomorphic Landform*
Debris flow	✗	✓	✓	✓	✗	✗	✓	✗	Debris deposit
Mass movement	✗	✗	?	✗	?	?	✗	✗	Landslide
Mudflow/slide	✗	✗	?	✓	?	?	✓	✗	Mudflow/mudslide
Solifluction	✗	✓	✓	✗	✗	✗	✓	✓	Solifluction lobe
Periglacial/permafrost	✗	✗	✓	✗	✓	✗	✓	?	Rock Glacier
Nival	✓	✗	✓	✓	?	✗	✓	?	Pronival rampart
Glacial	✓	✓	✓	✓	✓	✓	✓	✗	Moraine

and passive transport processes have taken place in the formation of these moraines, given the substantial quantity of fines, several bimodal distributions and the presence of subangular clasts. As discussed previously, bimodal particle size characteristics and fine material is typical of subglacially transported sediment, however the angular clasts reflect supraglacial transport. In addition, clasts in all trenches (with the exception of trench 3 for the eastern slope deposit) fall within the basal ice fabric envelope (Figures 7.17 and 7.20), once again suggesting active transport processes.

The western slope deposit displays attributes similar to lodgement till, subglacial melt out till and supraglacial melt out till (Table 2.6). It possesses several bimodal distributions in the three trenches (Table 7.1), the strong clast fabric in the direction of ice flow (Figure 7.16), similar isotropy and elongation (Figure 7.17) and well consolidated attributes (Figures 7.4 to 7.6) associated with lodgement till and subglacial melt out till. In contrast, this deposit also contains some unimodal distributions (Table 7.1) and a relatively high proportion of angular clasts (Figure 7.25 and Table 7.9), associated with supraglacial melt out till. Conversely, the eastern slope deposit contains fewer bimodal distributions than the western slope deposit (Table 7.4), and possesses a less significant proportion of fines (Figures 7.13 to 7.15 and Table 7.4). The random fabric characteristics in trenches 2 and 3 (Figure 7.19), together with predominantly angular clast shapes (Figure 7.29 and Table 7.14) are all typical of supraglacial melt out till.

These sedimentary results indicate that the Leqooa Valley deposits are moraines, which consist of both supraglacially and subglacially transported material in the case of the western slope deposit, whilst the eastern slope deposit is primarily composed of supraglacially transported material. This can be further quantified by comparing C_{40} index and roundness characteristics of the Leqooa Valley deposits to envelopes produced by Benn and Ballantyne (1994) on a covariance plot (Figure 9.9). The clasts sampled in the three trenches for both deposits fall within the moraine envelope. Results from the lateral moraines are similar to those for the Leqooa Valley deposits. The western slope deposit has C_{40} values ranging between 32 and 40 (Figure 7.22) and RA values between 34 and 50 (Figure 7.25 and Table 7.9), whilst the eastern slope deposit has C_{40} values ranging between 32 and 47 (Figure 7.26) and RA values

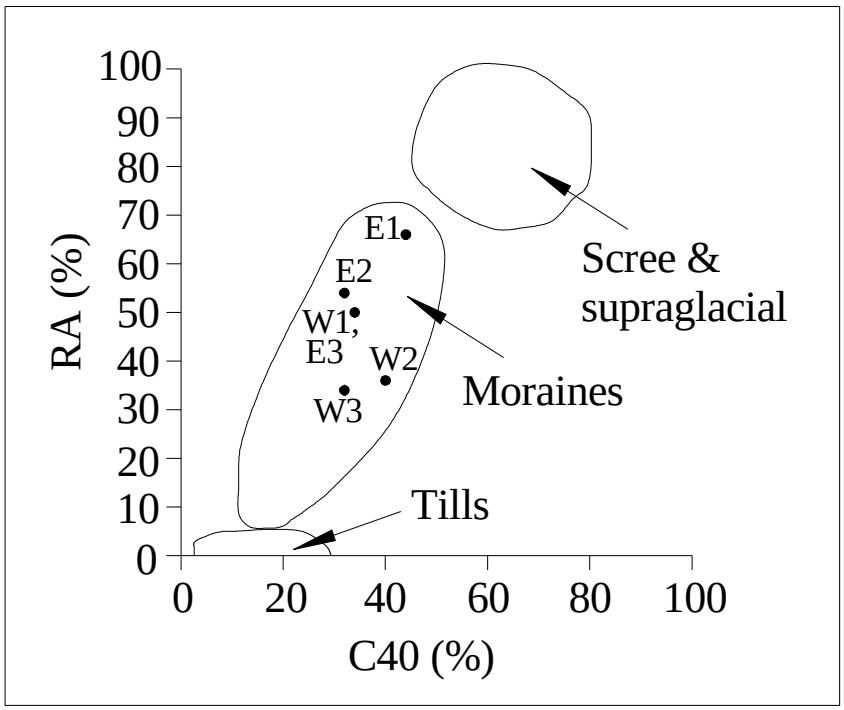


Figure 9.9 Co-variance plot of C_{40} and RA indexes and sediment envelopes (after Benn and Ballantyne, 1994) in comparison to the three trenches at the Leqooa Valley deposits (W1, W2, W3 = trenches 1, 2 and 3 for the western deposit, E1, E2 and E3 = trenches 1, 2 and 3 for the eastern deposit).

between 50 and 66 (Figure 7.29 and Table 7.14), both of which are similar to values described by Benn and Ballantyne (1994).

As stated previously, lateral moraine debris may be predominantly passively transported rockfall material which undergoes little modification, retaining its primary parent material characteristics which tend to be coarse, angular clasts with little matrix; however actively transported debris may also take place and result in matrix-rich debris in some dump moraines (Small, 1983; Benn and Evans, 1998; Huddart, 2004). The Leqooa Valley deposits display similar morphological and sedimentological characteristics to lateral moraines. The deposits provide evidence of passive transport in terms of particle size distribution, which is predominantly coarse-grained in the eastern deposit; however there is also evidence of active transport through the presence of finer material and subangular clasts in the western deposit. It is therefore suggested, based on the morphological and sedimentological characteristics that the deposits are lateral moraines.

9.12 POSSIBLE PROCESS ORIGIN FOR THE SEHONGHONG DEPOSITS

9.12.1 Deposits on the south-facing slope

A Summary of all morphological, sedimentological and micromorphological characteristics for the deposits located on the south-facing slopes of the Sehonghong River compared to various process mechanisms are presented in Table 9.6. The proposal that these deposits may be of mudflow or landslide origin may be dismissed on the basis that these deposits only contain the fabric characteristics of a mudflow, and none of the attributes of a landslide. There are several uncertainties regarding a fluvial origin for these deposits, given that clast shape and roundness is similar to that of fluvial deposits in the lower unit, yet not in the upper unit. However, there is a clear absence of stratification and preferred clast orientation in the lower unit deposits, therefore these are not considered to be of fluvial origin. The deposits possess all but the morphology and clast shape of debris flows, all but the clast shape of solifluction deposits and all but the morphology and micromorphology of glacial deposits. There is uncertainty regarding the micromorphological characteristics of the south-facing deposits with regards to debris flow processes, given that Site 1 possesses certain characteristics, whilst Site 2 does not. Although the morphology of these deposits does not concur with the typical morphologies of the associated process origins, these may still have contributed to the deposition of the Sehonghong deposits, without forming the distinct associated landforms. Clast shape may also be attributed to lithology, and cannot alone reject such origins. Although these deposits possess all the sedimentary characteristics of tills, these are highly variable deposits (Summerfield, 1991) and the lack of glacial landforms indicates that glacial processes did not operate at this site during the LGM. It is therefore suggested that the deposits occurring on the south-facing slopes of the Sehonghong River have been formed by both solifluction and debris flow processes.

As stated in Chapter 2, mass wasting occurs in the periglacial environment due to repeated freezing and thawing (Ballantyne and Harris, 1994). Frost susceptibility of sediment depends strongly on the amount of fines present (Konrad, 1999), with silts being most susceptible to the formation of segregated ice, as both suction potential and permeability are moderately high (Summerfield, 1991). The Sehonghong sediments are compared to a frost susceptibility diagram (Figure 9.10), which indicates that Site 1 at 45 cm depth and Site 2 at 50 cm depth host non frost

Table 9.6 Suitability of geomorphic processes and landform types for the various Sehonghong Sites 1 and 2 south-facing deposit attributes.

	Morph-ology	Particle Size	Sorting	Fabric	Eigen-value	Clast shape	Round-ness	Micro-morphology	
Geomorphic Process									*Geomorphic Landform*
Debris flow/avalanche	✗	✓	✓	✓	✓	✗	✓	?	Debris deposit
Landslide	✗	✗	?	✗	?	?	✗	✗	Landslide
Mudflow/slide	✗	✗	?	✓	?	?	✗	✗	Mudflow/mudslide
Solifluction	✓	✓	✓	✓	✓	✗	✓	✓	Solifluction/head deposit
Glacial	✗	✓	✓	✓	✓	✓	✓	✗	Till
Fluvial	✓	✗	✗	✗	?	?	?	?	Fluvial terrace

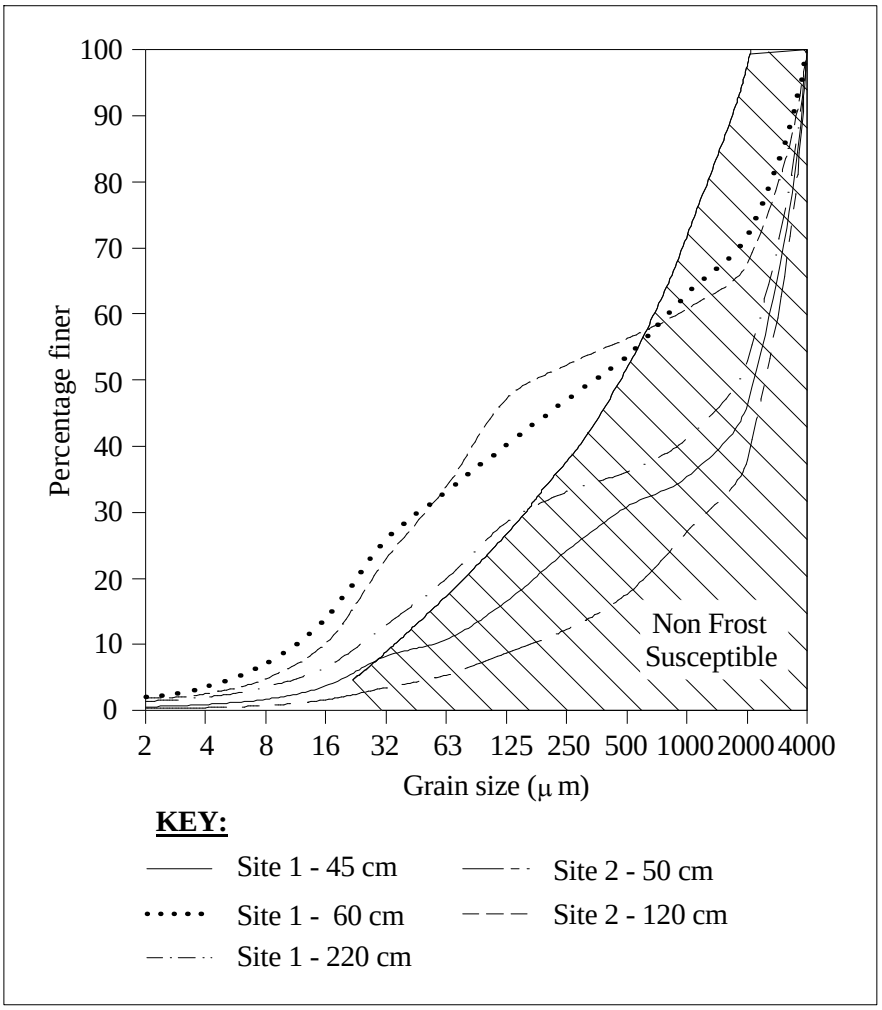


Figure 9.10 Grain-size curves for Sekhokong Site 1 and 2 compared to Beskow's (1935) frost susceptibility criteria (from Ballantyne and Harris, 1994).

susceptible sediments, whilst the finer material sampled at Site 1 at 60 cm and 220 cm depth and Site 2 at 120 cm depth are more frost susceptible, which suggests that these sediments would have been predisposed to the formation of segregated ice. This is further reinforced by the presence of silt cappings at these depths (Table 8.9), which suggests frost creep (Van Vliet-Lanoë, 1985b).

Ground temperatures were monitored at 3220 m a.s.l. in the Sehonghong area by Sumner (2003) and a potential freezing front at a maximum of 16 cm depth has been suggested. Diurnal freezing and thawing is only active in the upper 10 cm and most active in the first 5 cm, which indicates that the potential for sorted patterned ground is restricted to surficial sediments (Sumner, 2003). This corresponds to the typical size

of sorted stripes and circles found at these altitudes (Grab, 1996b). A 1ºC drop in temperature corresponds to an increase in maximum freezing depth of approximately 45 cm, therefore in order for freezing to penetrate to a depth of 1 m, a temperature depression of at least 2.5º C would be required (Sumner, 2003). Given a 6ºC decrease in temperature during the LGM, frost penetration could have exceeded 2 m, although permafrost is unlikely, due to the relatively high mean annual air temperatures (Boelhouwers, 1991; Sumner, 2003). In contrast, 50 cm frost depths were measured for a sorted circle and a stone-banked lobe near Mafadi (approximately 3400 m a.s.l.) in the central Drakensberg (Grab, 2004). The frost penetration rates were considerably slower for the sorted circle than for the stone-banked lobe, which has been attributed to the variable nature of solar radiation with regards to aspect, given that the sorted circle is on a north-facing slope, whilst the stone-banked lobe is on a south-facing slope (Grab, 2004). It has been suggested that contemporary frost penetration is likely to proceed beyond 50 cm depth for the stone-banked lobe and may even continue to at least 60 cm depth, based on the fact that the downward freeze component is well maintained down to 50 cm depth (Grab, 2004). These frost depths however occur beneath coarse material, which would allow for deeper frost penetration (Harris and Pedersen, 1998; Sumner, 2003).

Temperature data presented in Chapter 8 for Nhlangeni suggests that it is not cold enough for periglacial conditions to exist at the present time. However, during the LGM, temperatures are thought to have decreased between 5ºC and 7ºC (Heaton *et al.*, 1986; Talma and Vogel, 1992; Partridge *et al.*, 1999; Tyson and Partridge, 2000; Grab and Simpson, 2000), which indicates that annual air temperatures in the area would have been around -1.3ºC on the south-facing slope at 3200 m a.s.l. and 2.7ºC on the north-facing slope at 2650 m a.s.l., given a 6ºC drop, which places the area within the periglacial domain. The north- and south-facing slopes at 2650 m a.s.l. are at the boundary of the 3ºC periglacial limit suggested by French (1996), which indicates that this would have been the limit of periglacial activity in the Drakensberg during the LGM. Assuming an average global environmental lapse rate of 0.55ºC/100 m (Meyer, 1992), the contemporary mean annual summit temperature at Thabana Ntlenyana would be 3.2ºC, and hence temperatures during the LGM with a 6ºC temperature reduction would have been around -2.8ºC at this location, which concurs with findings by Grab (2002).

As would be expected, the south-facing slope is colder than the north-facing slope, particularly at 3200 m a.s.l. (Table 8.11), which suggests that periglacial processes would have been more active on these slopes. Snow cover on south-facing slopes is also more prominent than on north-facing slopes in the area (Grab, 1999b; Mulder and Grab, 2002), which would affect the amount of water available for solifluction and slopewash processes during spring (Harrison, 2002). The wide valleys also have gentle gradients and are less shaded against insolation, and are thus not conducive to the development of long-lasting snow patches, as opposed to incised valleys with steep slopes, which allow snow patches to develop (Harrison, 2002).

Colluvial mantles in the Lesotho Highlands dominate the steeper south-facing valley sides (Boelhouwers and Sumner, 2003). These mantles extend down to valley floors and are normally 2 m to 3 m in depth; however they can reach 15 m to 20 m in depth (Boelhouwers *et al.,* 2002). These mantles may show organically darkened a/b horizons and primarily constitute slope material (Sumner, 2003). Colluvial mantles which extend to the valley floors along the Sehonghong River are similar to those found in a few valleys to the south (Boelhouwers *et al.,* 2002). The colluvial sediments investigated by Sumner (2003) a few kilometres east of the Sehonghong site are made up of sand and gravel, whilst silt and clay make up less than 10%. Coarse gravel is most predominant in the upper 20 cm of the colluvial deposits and is considerably greater than that for the underlying layer (Sumner, 2003). This is very similar to the particle size characteristics of the Sehonghong Sites 1 and 2, which contain <11% silt and clay in the upper horizon, where gravel is the main constituent (Figures 8.9, 8.10 and Table 8.1). In addition, the debris mantles described by Boelhouwers *et al.* (2002) a few kilometres south of the Sehonghong deposits are very similar to those sampled at Sites 1 and 2, which contain a clastic horizon ranging between 20 cm and 70 cm in depth, underlain by a much finer sediment (Figures 8.6 and 8.7). It has been suggested that mass wasting probably took place during the LGM under conditions where snow cover was largely absent (Boelhouwers *et al.*, 2002).

It is suggested that the colluvial mantles which occur on the south-facing slopes along the Sehonghong River have formed through a number of periglacial slope processes operating both synergistically and on their own, which have subsequently created a

deposit, which represents a complex response to mass-wasting processes. It is further suggested that three distinct episodes of geomorphic activity have contributed to the formation of the colluvial mantles present on the south-facing slopes of the valley. Firstly, it is suggested that the lower units formed as a combination of mass wasting processes, including solifluction and debris flows, which is represented in the poorly sorted, angular debris, with downslope fabrics and where stratification is more or less absent. Secondly, it is suggested that the upper coarse gravelly layer was also formed as a result of mass wasting processes, however under different climatic conditions which enabled the mobilisation of larger clasts. Finally, a period of fluvial incision took place and exposed these sediments. The different process mechanisms for the formation of the two units are further reinforced in Figure 9.11, which indicates that the sediment at Site 2, sampled at 50 cm depth, falls outside the envelope for active solifluction sediments suggested by Harris (1987), whilst the sediment sampled at Site 1 at 45 cm depth falls just within the envelope. Alternatively, the sediments within the lower unit fall well within the solifluction envelope.

The different process mechanisms which formed these deposits would have operated under very different climatic conditions. It is suggested that the lower unit of finer sediment was deposited under periglacial conditions, enabling frost creep and solifluction processes to transport material downslope. This theory is reinforced by the presence of encircling silt cappings at 220 cm depth at Site 1 (Figure 8.19), which indicates that gelifluction processes dominated (c.f. Van Vliet-Lanoë, 1985b, 1987). These encircling cappings occur as a result of soil saturation from melting ice lenses and late-lying snow during the thaw phase (Elliot, 1996). In contrast, the surface cappings which occur at Site 2 at 100 cm depth are formed by frost creep where moisture supply is limited (Van Vliet-Lanoë, 1985b), once again indicating the varied nature of mass wasting processes which took place at this site. In order for the material in the upper coarse gravelly unit to be transported downslope, more powerful processes would have been necessary. It is suggested that more intense gelifluction processes would have been necessary to mobilise the size of material present in the upper gravelly unit, and hence implies deeper frost penetration associated with colder temperatures.

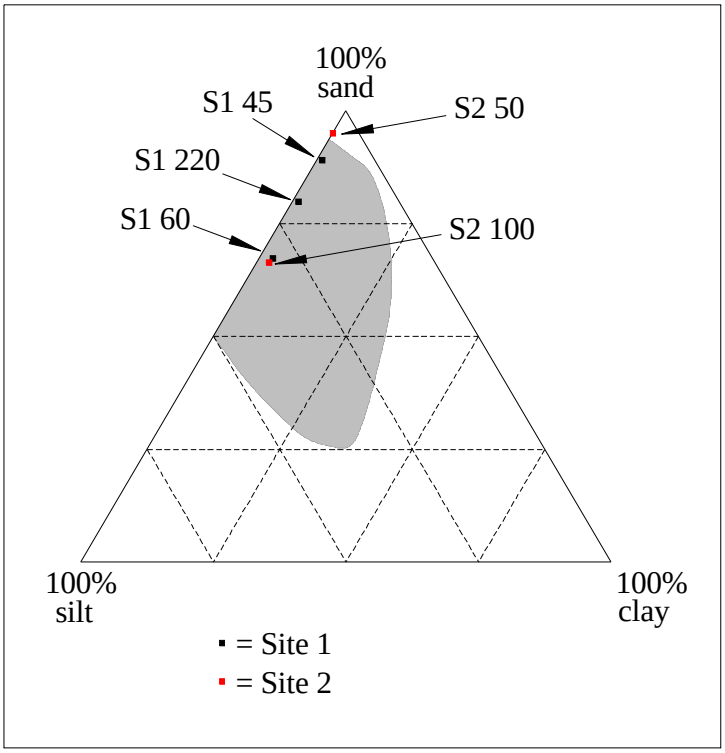


Figure 9.11 Ternary plot of sand-silt-clay for the matrix component of the Sehonghong Site 1 and 2 deposit compared to the textural envelope for active solifluction (after Harris, 1981a).

9.12.2 Deposits on the north-facing slope

In order to help identify a process origin for the deposit located on the north-facing slope of the Sehonghong River Valley, all morphological and sedimentological characteristics have been summarised in terms of whether they are typical attributes of the possible process mechanisms suggested for the formation of the deposit (Tables 9.7 and 9.8). An individual table has been devised for the upper and lower units at Site 3, as it is suggested from the sedimentological results that these may be the product of different processes. The upper units of coarse gravel possess only the fabric characteristics of mudflows, the morphological attributes of a fluvial terrace and none of the attributes of a landslide. However, these gravels possess all but the morphological attributes of debris flows and glacial deposits and all attributes typical to solifluction deposits. It is therefore suggested, based on the similarities between the gravelly units and the sedimentology of the deposits on the south-facing slope, that these are colluvial sediments, whilst the lower silt units represent a different process mechanism. The different processes involved in the deposition of the upper and lower units is further reinforced in Figure 9.12, where the matrix sampled from the upper

Table 9.7 Suitability of geomorphic processes and landform types for the Sehonghong Site 3 upper unit north-facing deposit attributes.

	Morph-ology	Particle Size	Sorting	Fabric	Eigen-value	Clast shape	Round-ness	
Geomorphic Process								*Geomorphic Landform*
Debris flow/avalanche	✗	✓	✓	✓	?	?	✓	Debris deposit
Landslide	✗	✗	?	✗	?	?	✗	Landslide
Mudflow/slide	✗	✗	?	✓	?	?	✗	Mudflow/mudslide
Solifluction	✓	✓	✓	?	?	?	✓	Solifluction/head deposit
Glacial	✗	✓	✓	✓	?	?	✓	Till
Fluvial	✓	✗	✗	✗	?	?	✗	Fluvial terrace

Table 9.8 Suitability of geomorphic processes and landform types for the Sehonghong Site 3 lower unit north-facing deposit attributes.

	Morph-ology	Particle Size	Sorting	Fabric	Eigen-value	Clast shape	Round-ness	
Geomorphic Process								*Geomorphic Landform*
Debris flow/avalanche	✗	✗	✓	✓	✗	✗	✗	Debris deposit
Landslide	✗	✗	?	✓	?	✗	✗	Landslide
Mudflow/slide	✗	✗	?	✗	?	✗	✗	Mudflow/mudslide
Solifluction	✓	✗	✓	✗	✗	✗	✗	Solifluction/head deposit
Glacial	✗	✓	✓	✓	✓	✓	✓	Till
Fluvial	✓	✓	✓	✗	?	✓	✓	Fluvial terrace

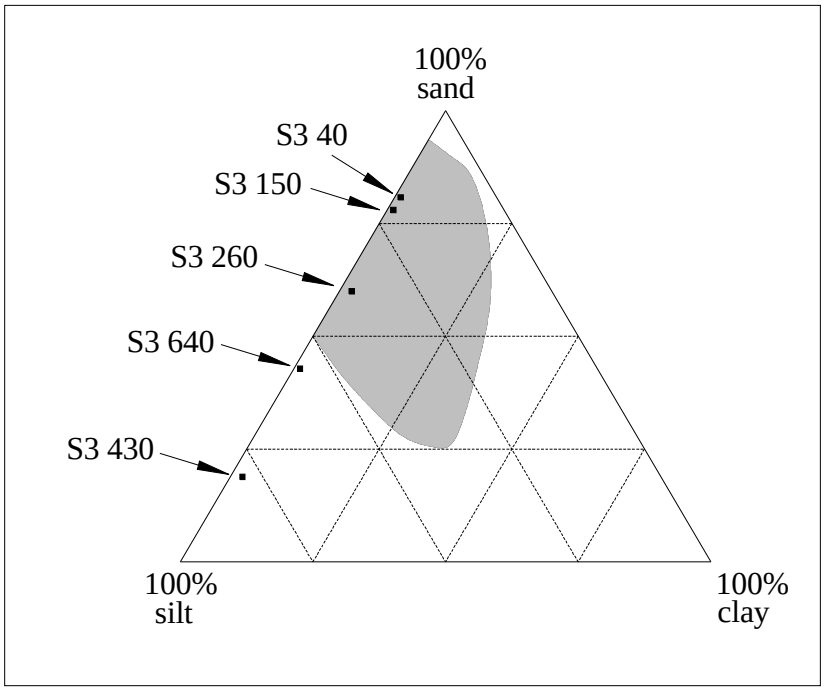


Figure 9.12 Ternary plot of sand-silt-clay for the matrix component of the Sehonghong Site 3 deposit compared to the textural envelope for active solifluction (after Harris, 1981a).

units fall within the solifluction envelope, whilst the sediment sampled within the silty unit falls outside it.

The lower unit does not possess the morphology, particle size, eigenvalue, shape or roundness of debris flows or any of the typical attributes of mudflows. The lower unit also only possesses the fabric characteristics of landslides and the morphology and sorting of solifluction deposits. In contrast, the lower unit possesses all but the morphological characteristics of tills and all but the fabric characteristics of fluvial deposits. The variability of tills may once again reflect the similarities between the glacial and fluvial characteristics. However, as stated previously, due to an absence of glacial landforms in the area, it is not envisaged that glaciers occupied the area during the LGM. It is therefore suggested that the lower silt units are of fluvial origin. However, the lack of upstream imbrication for these lower silt units suggests that this is not a channel deposit.

The lower unit of the Sehonghong Site 3 deposit contains >56% silt (Table 8.1), suggesting that processes capable of transporting larger material did not take place at the time of deposition in comparison to the south-facing deposits. It is suggested that the presence of two tributary streams in close proximity to the north-facing deposits played an important role in transporting sediment (Figure 9.13). Processes such as sheet flow/slope wash may have been initiated, possibly as a result of thin regolith cover during the summer months and the presence of a frozen layer during the winter months. Slope wash may be particularly effective in periglacial regions as the frozen soil prevents infiltration and favours surface run off (French 1996). It has been suggested in section 9.12.1 that areas above 2650 m a.s.l. would have fallen within the periglacial domain on both north- and south-facing slopes; therefore the seasonal freezing of the ground may have played an important role in the amount of surface runoff. Moisture for runoff may also be supplied by the melting of snow and melting of ice as the active layer thaws (French, 1996). The slightly higher temperatures experienced on the north-facing slopes may have increased diurnal thawing, which would have contributed greater amounts of moisture to the soil. Sheet flow sediments have been described as poorly sorted (Daniels, 2003), which is similar to sorting within the lower silt unit at Sehonghong Site 3 (Table 8.2).

The presence of gravel within the silt unit as indicated in Figure 8.8 suggests that processes were periodically sufficient to transport larger particles; however the mean a-axis of these clasts is only 4.7 cm, which is comparatively small. It is suggested that sheet flow transported the silts downslope, accumulating in the valley bottom where there would have been a relatively shallow accumulation of water. The clasts may then have rotated parallel to the slope as flow became more turbulent. This sheet flow would have preferentially been channelled in the tributary streams, thus maximising the transport of sediment downslope. This is further reinforced by the more rounded nature of the clasts sampled in the silt unit, which suggests abrasion and may well be attributed to fluvial processes. This area between the tributary deposits is also an area of preferential contemporary wetland development (Figure 9.13), further suggesting that these channels are an important source of groundwater. There is a lack of any pedogenic features within the sequence at Site 3, which suggests relatively high aggradation rates, which inhibit soil development (Daniels, 2003). It has been suggested that the threshold for aggradation rates that inhibit pedogenesis is about 0.5

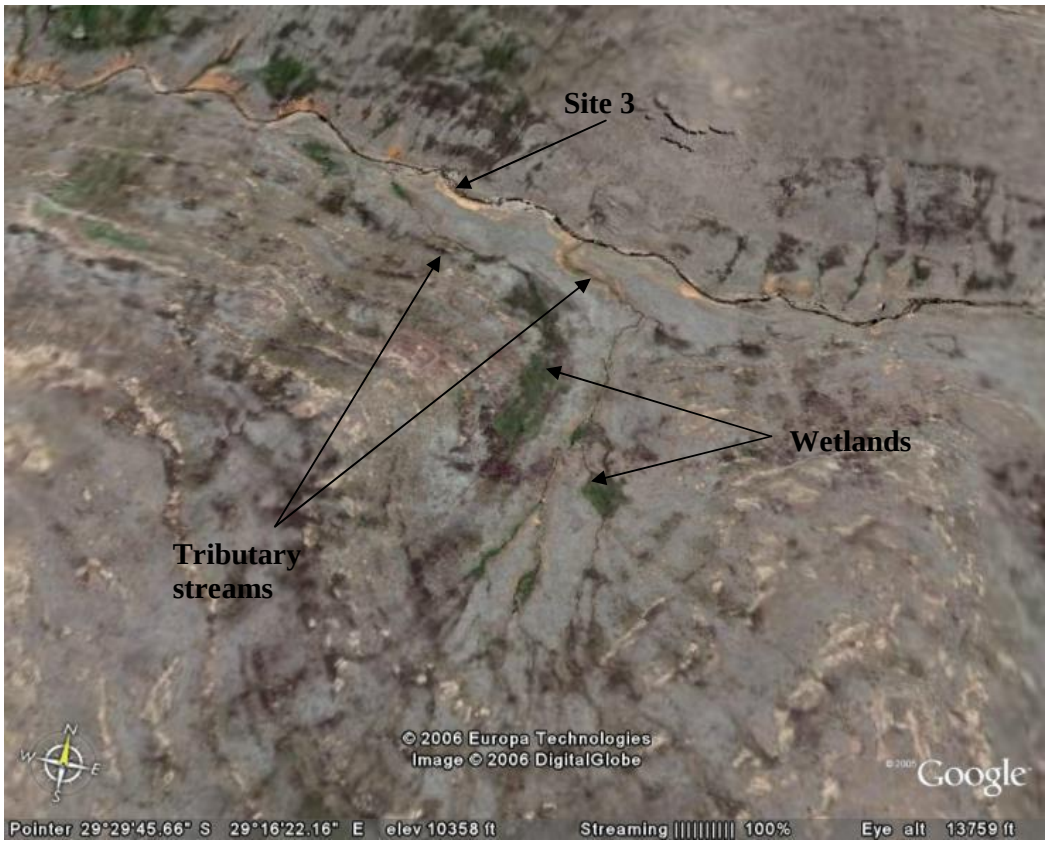


Figure 9.13 A view of the Sehonghong Site 3 (obtained from Google Earth).

cm year^{-1} (Daniels, 2003), implying that rates would have been greater than that at Sehonghong Site 3.

The difference between south- and north-facing slopes in South Africa has been described by Van Rheede van Oudtshoorn (1988) and Watson *et al.* (1984). Sediments on south-facing slopes are predominantly unconsolidated, particle size ranges from large boulders to fine sand and gravel and these sediments are highly erodible once the protective vegetation cover has been removed (Van Rheede van Oudtshoorn, 1988), whilst the north-facing slopes consist of colluvial deposits containing generally finer material (<2mm) (Watson *et al.,* 1984). This concurs with the findings in this research, where the south-facing slope deposits consist of gravel with large boulders, whilst the north-facing slope sediments also contain high percentages of silt (Table 8.1). The mobilisation of these gravels on a north-facing slope as a result of increased run off corresponds with suggestions made for the process origins for the north-facing Sehonghong deposits. Further sedimentary

sequences at Golden Gate have been described as valley fill deposits incised by the Little Caledon River, consisting of unsorted matrix-supported diamicts, clast-supported diamicts with subangular clasts up to 1 m in length, fluvially reworked deposits laid as gravel and sand beds and palaeosols (Marker, 1994b). It has been suggested that these diamicts were deposited during the LGM, based on dates obtained from palaeosols underlying these diamicts (Marker, 1995). Further evidence of clastic sediments overlying organic sediments has been observed at Tlaeeng Pass in Lesotho on north- and northwest-facing slopes (Hanvey and Marker, 1994). These occur in hollows and have subsequently been incised by fluvial processes (Hanvey and Marker, 1994). It is suggested that the diamicts are similar to solifluction material and that the sediments originated from slope weathering and were subsequently transported downslope by cryogenic processes (Hanvey and Marker, 1994). These matrix-supported and clast-supported diamicts are very similar to those described for both the south- and north-facing slopes along the Sehonghong River.

9.13 RADIOCARBON DATING

Material tested for radiocarbon dating was soil organic matter (SOM), which is generally introduced into the soil as either root material or organic detritus at the surface. There were very small amounts of SOM in the samples, suggesting a slow biological turnover. The calibrated ages for the five linear deposits range between 13 820 yrs BP and 20 530 yrs BP (Table 9.9). There is a fairly wide range in ages within the Tsatsa-La-Mangaung deposit and between deposits; however the younger dates may reflect the intrusion of slightly younger soil organic matter at the depths sampled, or older material which accumulated prior to deposition. These ages probably significantly postdate the deposition of the deposits, given that they represent soil which was picked up at the front of the glacier and subsequently deposited within the moraine. They indicate that the moraines were formed towards the end of the Last glacial period after 13 820 yrs BP.

Recent work by Holgrem *et al.* (2003) from a stalagmite in the Makapansgat Valley suggests that although maximum cooling occurred around 17 500 yrs BP in southern Africa, the time period between 15 000 yrs BP to 13 500 yrs BP was also a cold period and renewed warming occurred after about 13 500 yrs BP. It is possible that the secondary ridges at Tsatsa-La-Mangaung (Figures 5.1 and 5.2), Sekhokong Site 1

Table 9.9 A summary table of the radiocarbon and calibrated ages for the study sites.

Sites	Radiocarbon Age yrs BP	Calibrated Age yrs BP
Tsatsa-La-Mangaung	13 790 ± 110	16 500
	17 300 ± 100	20 530
	11 700 ± 70	13 820
Sekhokong Site 1	12370 ± 60	14 700
Sekhokong Site 2	16270 ± 70	19 350
Leqooa western deposit	14 400 ± 60	17 200
Leqooa eastern deposit	12 890 ± 60	15 390
Sehonghong	19 800 ± 580	23 055
	25 900 ± 670	27 410
	36 600 ± 1400	43 085

(Figures 6.1 and 6.2) and Leqooa Valley (Figures 7.2 and 7.3) may reflect these two cold periods. It is possible that these secondary ridges reflect the former extent of palaeoglaciers, which were however not well developed due to limited sediment supply, which only became available towards the end of a relatively short glacial period.

Further evidence for a cold period prior to 13 500 yrs BP has been proposed by Marker (1994a). Sedimentary sequences have been described from a north-facing slope near Sani Top and reveal organic sediments overlying diamicts or gravels (Marker, 1994a). The organic sediments yielded dates between 1000 ± 40 yrs BP and 13 490 ± 130 yrs BP, and the change from gravels to peat in the sedimentary sequences has been attributed to a change in sedimentation from inorganic to organic (Marker, 1994a). It is suggested that the thick gravel horizons indicate cold, dry conditions, which prevented organic accretion and enabled higher-intensity runoff as a result of reduced vegetation cover, which in turn led to increased slope movement (Marker, 1994a, 1995). It was therefore suggested that the period prior to 13 500 yrs BP was colder, encouraging the mobilization of the gravels, whilst the period 13 500 yrs BP to 9000 yrs BP was warmer and wetter, as indicated by a phase of organic sedimentation (Marker, 1994a).

The ages obtained for the deposits also correspond well with documented glacial advances in Southern Hemisphere regions such as the South American Andes (Mercer, 1972, 1976; Porter, 1981a; Denton *et al.*, 1999), Mount Kenya (Mahaney, 1987; Mahaney *et al.*, 1991, 2000), and the New Zealand Alps (Moar, 1980; Denton *et al.*, 1999; Hellstrom *et al.*, 1998; Benn and Evans, 1998; Kirkbride and Brazier, 1998). Final deglaciation in the South American Andes began after 11 000 yrs BP (Mercer, 1976; Heusser, 1993; Heusser *et al.*, 1999; McCulloch *et al.*, 2000), after 12 000 yrs BP in the New Zealand Alps (Chinn, 1983; Hellstrom *et al.*, 1998), and after 12 500 yrs BP on Mount Kenya, East Africa, where moraines have been dated at 12 500 yrs BP (Mahaney *et al.*, 1991).

The radiocarbon dates obtained by Prof. Stefan Grab at Sehonghong Site 3 (Table 8.10) range from 23 055 yrs BP to 43 085 yrs BP. These dates represent a minimum age, and the oldest age may also represent an infinite age beyond the 40 000 year limit of ^{14}C dating. The unit that has been identified at 8.5 m depth and dated at 43 085 yrs BP is attributed to a relatively wet period with wetland development (Grab *et al.* in prep). An overlying gravely unit indicates a drier period, with a possible renewed phase of wetland development at 27 410 yrs BP (Table 8.10). It is then suggested that fluvial sedimentation occurred before a renewed wetland development period at 23 055 yrs BP (Grab *et al.* in prep), which concurs with a more humid phase suggested between 25 000 yrs BP and 21 000 yrs BP in southern Africa (Tyson, 1986). Although the ages are in stratigraphic order, their time of deposition cannot necessarily be attributed to a specific depositional period being cold or warm as the organic matter tested may be older than the time at which it was deposited, nevertheless, the alternating units of organic sand and gravel are believed to indicate wetter or drier periods (Grab *et al.* in prep). In previous studies, the presence of coarse-grained with fine-grained organic rich sediments has been attributed to episodic high-energy sediment transport events (Goodbred and Hine, 1995; Goodbred *et al.*, 1998; Mitsch and Gosselink, 1986). It is therefore problematic to try and relate the alternating units of sand and gravel to past climatic events.

Radiocarbon ages obtained from colluvial deposits in Swaziland range between 31 420 ± 780 yrs BP and 11 660 ± 50 yrs BP, which indicates that colluviation

commenced after 40 000 yrs BP, whilst it terminated between 12 000 yrs BP and 10 000 yrs BP (Watson *et al.*, 1984), which corresponds with the ages obtained for the Sehonghong Site 3 deposit (Table 9.1). It has been suggested that colluviation is initiated by the onset of drier conditions, which causes a reduction in vegetation cover, therefore allowing the movement of previously weathered material (Watson *et al.*, 1984). This indicates a drier climate than at present during the deposition of colluvial material on both the south- and north-facing slopes at Sehonghong. It has further been suggested that the most obvious period for a drier climate would have been around the LGM (Watson *et al.*, 1984).

9.14 SUMMARY

The linear slope deposits at Tsatsa-La-Mangaung, Sekhokong and Leqooa Valley have been attributed to glacial processes based on their morphological, sedimentological and micromorphological characteristics. It is suggested that the Tsatsa-La-Mangaung, Sekhokong Site 1 and Leqooa Valley deposits are marginal moraines, whilst the Sekhokong Site 2 deposit is a debris flow deposit which formed as a result of melting of a small glacier. In contrast, the Sehonghong deposits have been attributed to various mass wasting processes formed under a various climatic regimes. These deposits, along with the considerable literature available on colluvial and periglacial deposits in the Drakensberg, indicate that the environment during the LGM in the high Drakensberg was predominantly periglacial. The general consensus is that South Africa was never glaciated during the Quaternary (Tyson, 1986; Preston-Whyte and Tyson, 1988; Tyson and Preston-Whyte, 2000; Deacon and Lancaster, 1988; de Villiers, 2000); however, the linear deposits identified as moraines in four localities indicate that small glaciers did exist in this region, forming moraines dated back to the late glacial period. It is not suggested that widespread glaciation occurred in the Drakensberg during the LGM, rather that glaciers formed a minor component within a periglacial land system.

CHAPTER 10

Glacier reconstruction

10.1 INTRODUCTION

It has been suggested in Chapter 9 that the deposits at Tsatsa-La-Mangaung, Sekhokong Site 1 and Leqooa Valley are lateral moraines formed by small glaciers, whilst the deposit at Sekhokong Site 2 is a debris flow deposit, which was initiated by the melting of a small glacier. In order to quantify such hypotheses, it is possible to model the palaeo-reconstruction of former glaciers, and further verify the identification of these features. This glacier reconstruction work will permit the differentiation of whether the deposits were formed in a passive way by snowbanks, or actively by glaciers (Carr, 2001).

10.2 CLIMATE DURING THE LAST GLACIAL MAXIMUM

Temperature data have been recorded for Sani Top (2878 m a.s.l.), which is approximately 2 km from the Tsatsa-La-Mangaung site (Table 10.1). Unfortunately the data loggers at the Leqooa Valley site were stolen and thus extrapolations will be made from the Sani top data. Only the summer months are taken into account (December, January and February for the Southern Hemisphere) (c.f. Ohmura *et al.*, 1992), because there is a close, non-linear relationship between accumulation and mean summer temperature at the equilibrium line (Humlum, 1997). Data for the period 2002 to 2003 and 2004 to 2005 are missing due to faulty data loggers. The average temperature for the four available years is 10ºC and it should be noted that the temperature appears to be rising over the years. Temperature records for Shaleburn (29º48’0”S, 29°21’0”E at 1614 m a.s.l.), which is the closest station to Sani Top are provided in Figure 10.1. These records illustrate that the years 2000/2001 and 2001/2002 were below the 20 year average, whilst the four subsequent years were above average. This suggests that the temperature records for Sani Top are a reasonably accurate representation of the average temperature during the logging period.

Table 10.1 Summer temperature data for Sani Top.

	2000-2001	2001-2002	2003-2004	2005-2006
December	10.35	9.30	10.23	9.53
January	10.00	10.86	10.12	11.00
February	8.93	9.27	9.67	10.54
Average	**9.76**	**9.81**	**10.01**	**10.36**

Some of the first ridges to be discovered in the high Drakensberg and attributed to glacial processes were located at Nhlangeni and described by Grab (1996a). The average yearly temperature data at this site have been presented in Chapter 8, however mean summer temperature is presented here for the north and south-facing slope in order to compare with the linear deposit sites (Table 10.2 and 10.3). Temperatures at 2900 m a.s.l. on the south-facing slope average 10.7ºC for the period 1999 to 2001, whilst they average 10.8ºC at the same elevation on the north-facing slope. These temperatures compare well with the data from Sani Top. At an altitude of 3200 m a.s.l. on the south-facing slope, the mean summer temperature for the period 1999 to 2003 is 9ºC, which is 1.7ºC colder than at 2900 m a.s.l., indicating a lapse rate of 0.57ºC/100 m. This compares well to the lapse rate of 0.55ºC/100 m suggested by Meyer (1992). In contrast, on the north-facing slope, the difference in average temperature between 2900 m a.s.l. and 3200 m a.s.l. is 2.3ºC, indicating a lapse rate of 0.77ºC/100 m. There was a substantial difference in mean annual air temperatures between the north- and south-facing slope as indicated in Chapter 9, where the north-facing slope was warmer. However, the mean summer temperatures indicate that there is no real variation between the north- and south-facing slopes at Nhlangeni. The mean summer temperatures at Nhlangeni may be a slight underestimation as the recording time period was below average compared to the 20 year mean (Figure 10.1).

Southern African temperature decreases during the LGM are thought to have been between 5ºC and 7ºC (Heaton *et al.,* 1986; Talma and Vogel, 1992; Partridge *et al.,* 1999; Tyson and Partridge, 2000; Grab and Simpson, 2000; Holmgren *et al.,* 2003), whilst precipitation along the Drakensberg was reduced by approximately 30% (Partridge *et al.,* 1999; Tyson and Partridge, 2000). Elsewhere in the Southern Hemisphere, LGM temperatures were depressed by 6ºC to 8ºC (Barnola *et al.,* 1987; Denton *et al.,* 1999), and thus the suggested southern African decreases are not exceptional.

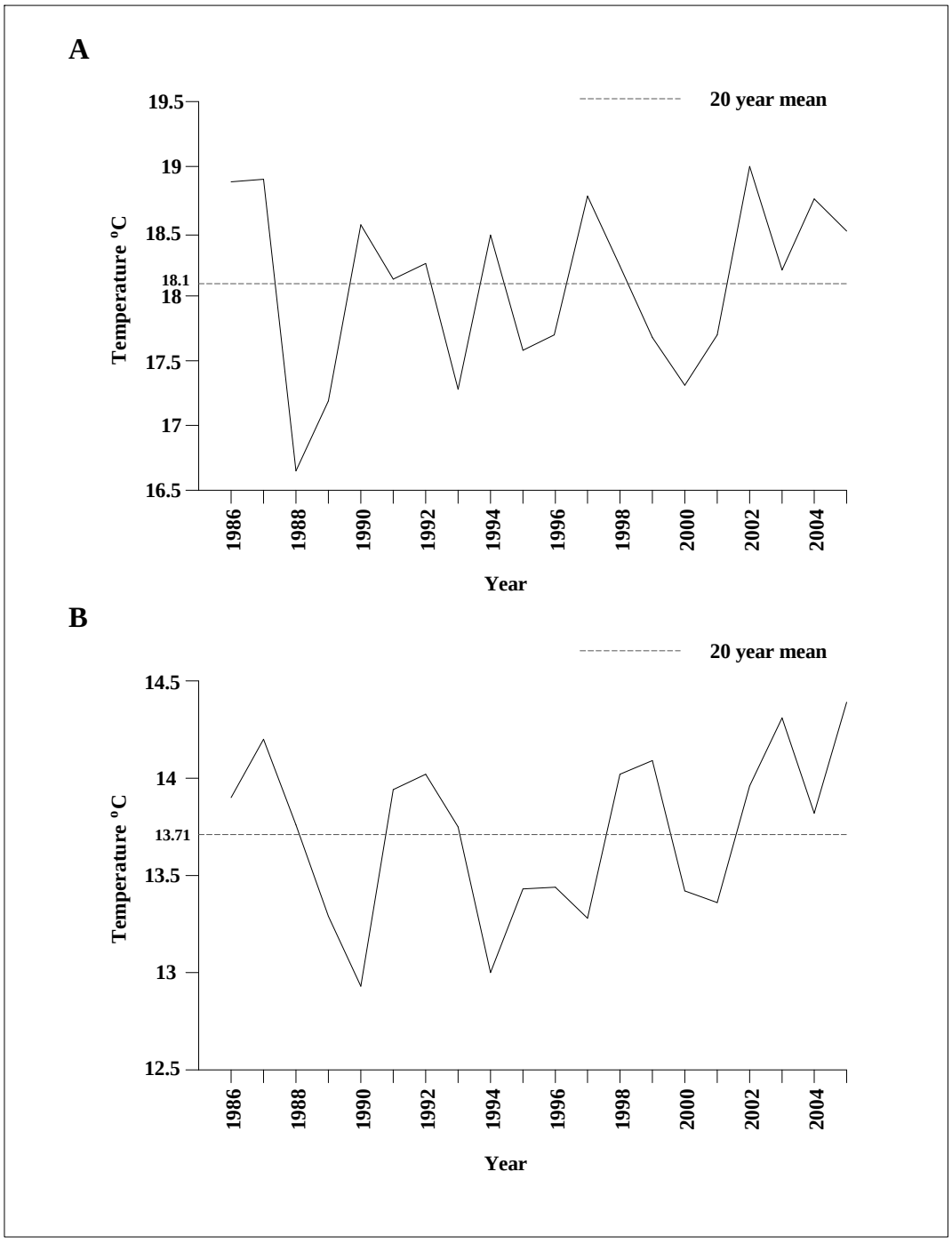


Figure 10.1 A 20 year temperature record for Shaleburn. A = Mean annual summer temperature (December to February), B = Average yearly temperature.

Table 10.2 Summer temperature data for Nhlangeni south-facing slope.

Site	Month	1999-2000	2000-2001	2001-2002	2002-2003
2650 m a.s.l	December	11.93	12.25	-	-
	January	11.09	12.18	-	-
	February	12.62	11.33	-	-
	Average	**11.88**	**11.92**	**-**	**-**
2900 m a.s.l.	December	10.57	11.28	-	-
	January	9.76	11.17	-	-
	February	11.20	10.17	-	-
	Average	**10.51**	**10.87**	**-**	**-**
3200 m a.s.l.	December	8.41	9.07	8.37	8.51
	January	8.01	9.38	9.97	10.29
	February	9.30	8.00	8.70	10.27
	Average	**8.57**	**8.82**	**9.01**	**9.69**

Table 10.3 Summer temperature data for Nhlangeni north-facing slope.

Site	Month	1999-2000	2000-2001	2001-2002	2002-2003
2650 m a.s.l	December	11.76	12.39	-	-
	January	10.80	11.90	-	-
	February	12.26	10.97	-	-
	Average	**11.61**	**11.75**	**-**	**-**
2900 m a.s.l.	December	10.64	11.14	-	-
	January	9.96	11.15	-	-
	February	11.58	10.29	-	-
	Average	**10.72**	**10.86**	**-**	**-**
3200 m a.s.l.	December	8.12	8.96	-	8.45
	January	7.65	9.47	-	10.32
	February	9.49	8.37	-	10.47
	Average	**8.42**	**8.93**	**-**	**9.75**

An average temperature decrease of 6ºC is assumed for the purpose of this research; therefore mean temperatures at Sani Top would have been 4ºC during the LGM. The mean summer temperature at Nhlangeni would have been 5.9ºC and 5.7ºC at 2650 m a.s.l. for the south- and north-facing slope respectively, 4.7ºC and 4.8ºC at 2900 m a.s.l. for the south- and north-facing slopes and 3ºC at 3200 m a.s.l. for both the south- and north-facing slope. Assuming an average global environmental lapse rate of 0.55ºC/100 m (Meyer, 1992), it is also possible to calculate the temperature at the ELA for each linear deposit site (see further sections for ELA calculations).

Contemporary meteorological data in the Drakensberg and Lesotho are scarce (Boelhouwers and Meiklejohn, 2002), and measured rainfall for the Drakensberg escarpment above 2500 m a.s.l. has, until recently, not existed on record (Nel and Sumner, 2005). All rainfall data for the high Drakensberg obtained in the past are derived by projection from stations at lower altitudes (Tyson *et al.,* 1976; Schulze, 1979) and there have been no records in recent years to verify these estimates (Nel and Sumner, 2005). It has been suggested that contemporary precipitation amounts to between 1500 mm/annum and 1600 mm/annum along the escarpment (Killick, 1963; Schulze, 1979). However, more recently precipitation records for Sani top have been recorded, which suggest that this area receives between 742 mm/annum (Nel and Sumner, 2005) and 821 mm/annum (Sene *et al.,* 1998). These more recent lower precipitation records are regional in the case of Sene *et al.* (1998) and only limited to one year (2002) in the case of Nel and Sumner (2005). This year was particularly drier than average and therefore may not be a true representation of average precipitation (Figure 10.2). Furthermore, the data provided by Nel and Sumner (2005) do not record snowfalls and there are no snowfall records available for the escarpment area (Nel and Sumner, 2005). Inaccuracies in the measurement of rainfall have also been discussed by Nel and Sumner (2005), notably the rain-catch deficiency caused by windy conditions. It has been reported that rain-catch deficiencies are in excess of 8.1% (Schulze, 1979) and may be as much as 20% for windy sites (Rodda, 1967; Schulze, 1979). It is for this reason that an average has been taken from Killick (1963), Schulze (1979), Sene *et al.* (1998) and Nel and Sumner (2005), which amounts to 1170 mm/annum. If precipitation was reduced by 30% during the LGM (Partridge *et al.,* 1999; Tyson and Partridge, 2000), then precipitation at Sani Top could have averaged ca. 820 mm/annum.

Having postulated palaeotemperature and palaeoprecipitation data for the Tsatsa-La-Mangaung, Sekhokong and Leqooa Valley sites, it is possible to compare these to present-day glaciated regions. Ohmura *et al.* (1992) produced a best-fit climate curve for the ELA of 70 glaciers, which offers opportunities to make deductions on the likely climatic changes required for glaciation at the Drakensberg study sites. Given the palaeoclimatic extrapolations for the Drakensberg, it is possible to plot the likely

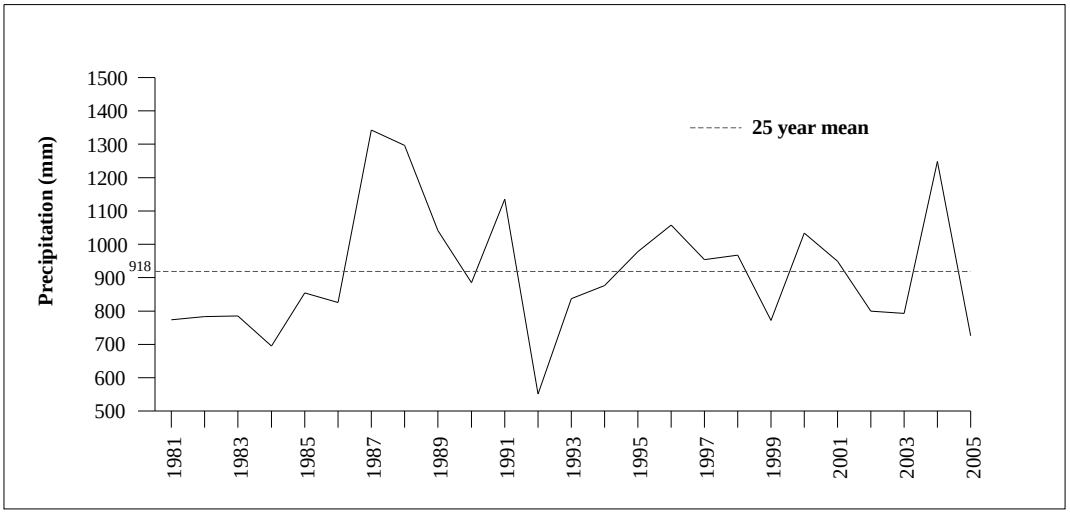


Figure 10.2 A 25 year precipitation record for Himeville (29º45'0"S, 29º30'0"E at 1579 m a.s.l.), which is the closest recording station to the study sites.

LGM summer temperature and annual precipitation for the Tsatsa-La-Mangaung, Sekhokong and Leqooa Valley sites on the ELA curves produced by Ohmura *et al.* (1992) (Figure 10.3).

The red triangles indicate that the four study sites fall just outside the best-fit curves for glacier ELAs, assuming a 6ºC decrease in temperature and a 30% decrease in precipitation. The black triangles, on the other hand, indicate the necessary precipitation required for glaciers to be sustained given a 6ºC temperature depression, which is significantly higher than recorded precipitation data. Nevertheless, either a small decrease in temperature, or a small increase in precipitation would place these sites within the boundaries of the best-fit curves. A 7ºC decrease in temperature would satisfy the winter accumulation at the ELA, whilst a total of between 1400 mm and 1600 mm precipitation would be required to sustain glaciers given a 6ºC temperature depression.

Palaeotemperature and precipitation data for Nhlangeni have also been plotted onto the Ohmura graph in order to give an indication of the difference that altitude and aspect makes with regards to the feasibility of the presence of glaciers at this site. As stated previously, aspect does not affect mean summer temperature at this site, which

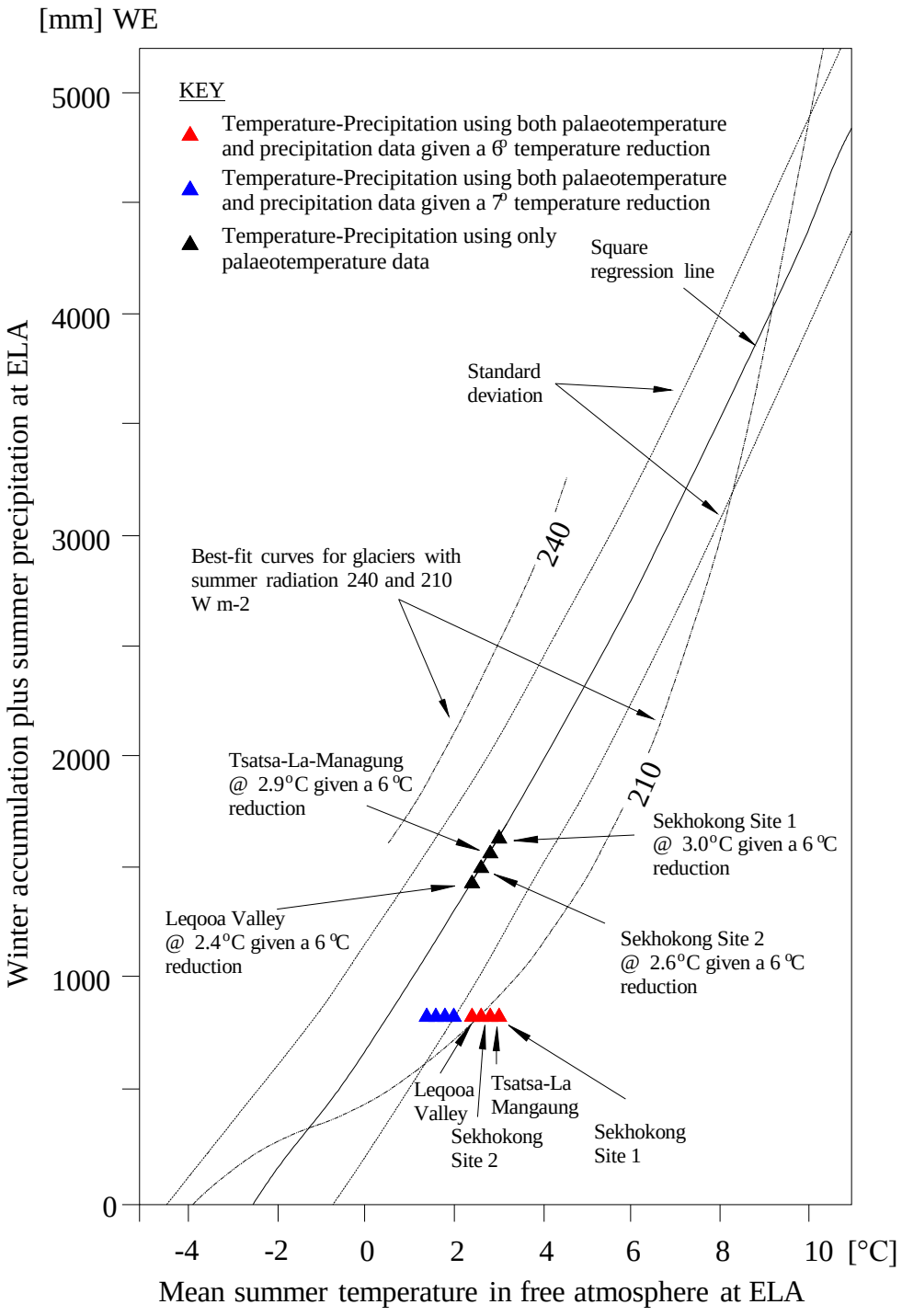


Figure 10.3 Plotting of the high Drakensberg LGM palaeoclimates for the Tsatsa-La-Mangaung, Sekhokong and Leqooa Valley sites, with the best-fit climate curve for the ELA of 70 glaciers (after Ohmura *et al.*, 1992).

is clear from Figure 10.4 and 10.5, as these are almost identical. In contrast, the effect of altitude on temperature, and hence the best fit curve of glaciers is clear. Only at 3200 m a.s.l. is Nhlangeni close to the best fit curve, and a 7ºC temperature drop would place this site within the envelope for glaciers. In contrast, an 8.7ºC and 9.9ºC drop in temperature would be required at 2900 m a.s.l. and 2650 m a.s.l. respectively. The required precipitation values are also somewhat unrealistic, given that with a 6ºC drop in temperature at the LGM, precipitation would have had to have amounted to 2200 mm and 2700 mm at 2900 m a.s.l. and 2650 m a.s.l. respectively. It is therefore suggested that should glaciers have existed, based on the 6ºC drop indicated by proxy data, these would most likely have been at sustained altitudes in excess of 3000 m a.s.l.

Mean summer temperature on modern glaciers is reported to be closely related to winter accumulation at the ELA (Sutherland, 1984; Ballantyne, 1989b; Ohmura *et al.*, 1992; Lie *et al.*, 2003), and it is suggested that proxy evidence in the use of estimating palaeotemperature is significant when identifying snow accumulation (Ballantyne, 2002). In addition, Bakke *et al.* (2005) indicate that the use of temperature to derive former winter precipitation is beneficial in reconstructing former glacier fluctuations. The temperature data collected at Sani Top are believed to be more reliable than the precipitation data. Temperature depreciation during the LGM is based on analyses of dissolved gasses from fossil aquifers (Heaton *et al.*, 1986) and speleothems (Talma and Vogel, 1992; Holgrem *et al.*, 2003) and is believed to be more robust than the estimated isolines of mean precipitation suggested by Partridge (1997), given the marked effects of the escarpment. It has been suggested that using temperature to estimate palaeoprecipitation is significant in reconstructing glacier mass balance (Bakke *et al.*, 2005). The diagram devised by Ohmura *et al.* (1992) summarises the most reliable relationship between temperature and mass loss at the ELA (Carr and Coleman, 2006) and therefore the derived temperature values during the LGM indicate the likely precipitation at the ELA for the four study sites.

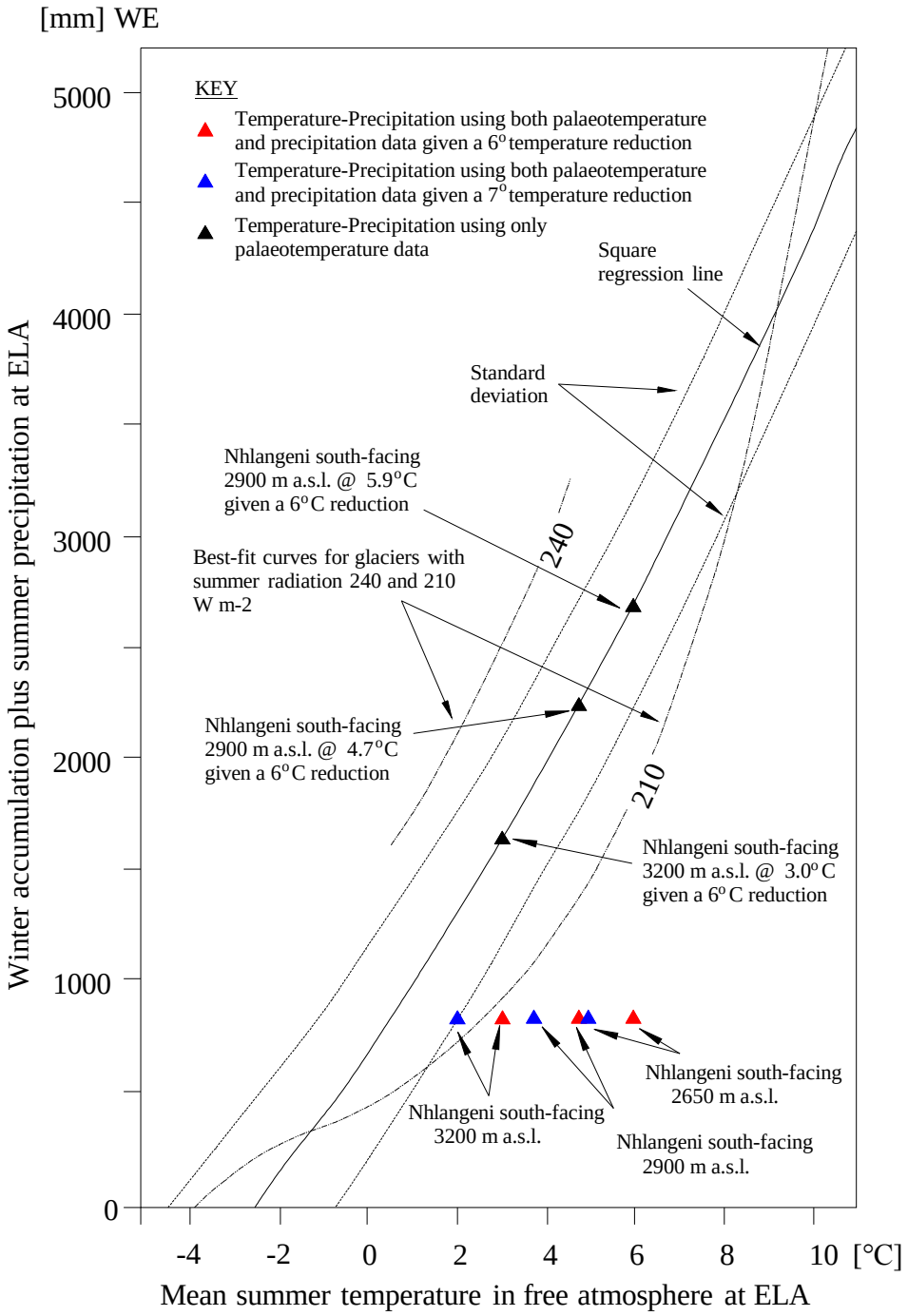


Figure 10.4 Plotting of the high Drakensberg LGM palaeoclimates for the Nhlangeni south-facing slope, with the best-fit climate curve for the ELA of 70 glaciers (after Ohmura *et al.*, 1992).

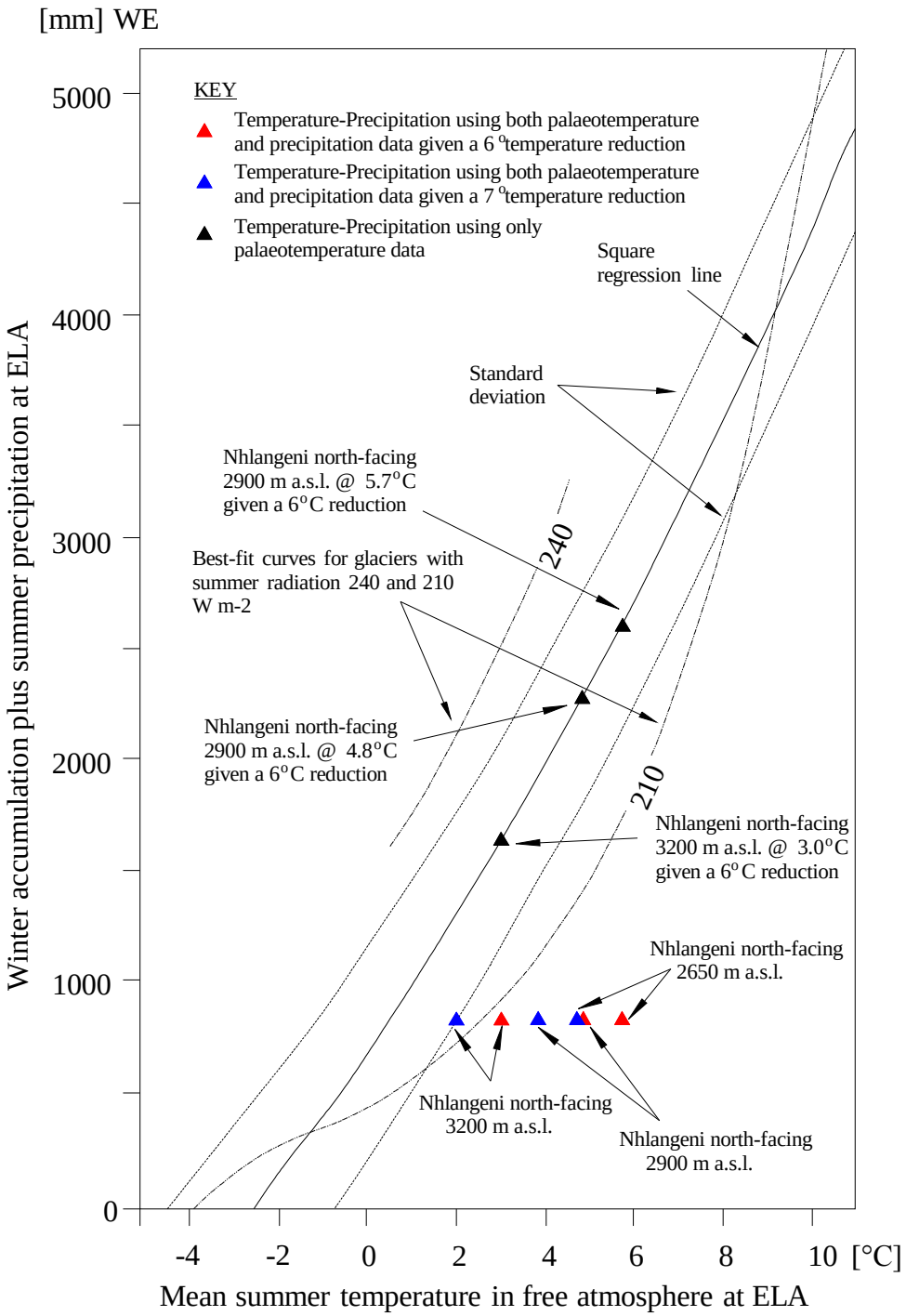


Figure 10.5 Plotting of the high Drakensberg LGM palaeoclimates for the Nhlangeni north-facing slope, with the best-fit climate curve for the ELA of 70 glaciers (after Ohmura *et al.*, 1992).

10.3 GLACIER RECONSTRUCTION

The reconstructed palaeoglaciers at Tsatsa-La-Mangaung, Sekhokong and Leqooa Valley are presented in Figures 10.6 to 10.9. The calculated ablation gradient based on the best-fit curve of Andrews (1972) for each glacier is represented in Table 10.4, whilst the mass balance characteristics of the palaeoglaciers are presented in Table 10.5. The ELA of the former glaciers vary through an altitudinal range of 88 m, whilst the ablation gradients vary from 4.6 mm m^{-1} at Leqooa Valley to 4.9 mm m^{-1} at Tsatsa-La-Mangaung and Sekhokong Site 1. These moderate ablation gradients reflect glacier activity in relatively mild maritime conditions (Schytt, 1967). The average total velocity at the ELA ranges from 1.15 m a^{-1} at Tsatsa-La-Mangaung to 4.5 m a^{-1} at the Leqooa Valley site. This is comparable with modern analogues, where most glacier velocities range between 1 m a^{-1} to 100 m a^{-1}, with the exception of surging glaciers and very rapid outlet glaciers (Andrews, 1972). Basal shear stress ranges from 0.50 bars at Sekhokong Site 1 to 0.64 bars at Leqooa Valley, which again falls within the range of modern analogues. Basal shear stress in alpine valley glaciers is usually between 50 kPa and 150 kPa (100kPa = 1 bar) (Nye, 1952; Murray and Locke, 1989; Paterson, 1994; Menzies, 2002a). Further work on basal shear stress suggests that glacial reconstructions should obey the same physical laws as glaciers today and should therefore have basal shear stress values of between about 50 kPa and 100 kPa over hard substrates (Bennett and Glasser, 1996; Nesje and Dahl, 2000). Shear stresses below 1 bar produce very small rates of strain, whilst above this, they produce larger rates of deformation than those that exist in glaciers and ice sheets (Nye, 1952).

Movement within temperate glaciers is predominantly by basal slip and internal deformation (Dahl *et al.,* 1997). In ice deformation, the ice deforms under its own weight to relieve the internal stresses, and may be described by Glen's flow law (Van der Veen, 1999). Basal sliding becomes important where basal temperatures have reached the pressure melting temperature and subglacial water is present (Van der Veen, 1999). Basal slip has been measured to account for 10% to 90% of total velocity (Andrews, 1975; Paterson, 1981) and will be at a minimum where basal shear stress is at its maximum (Murray and Locke, 1989). More recently, Carr and Coleman

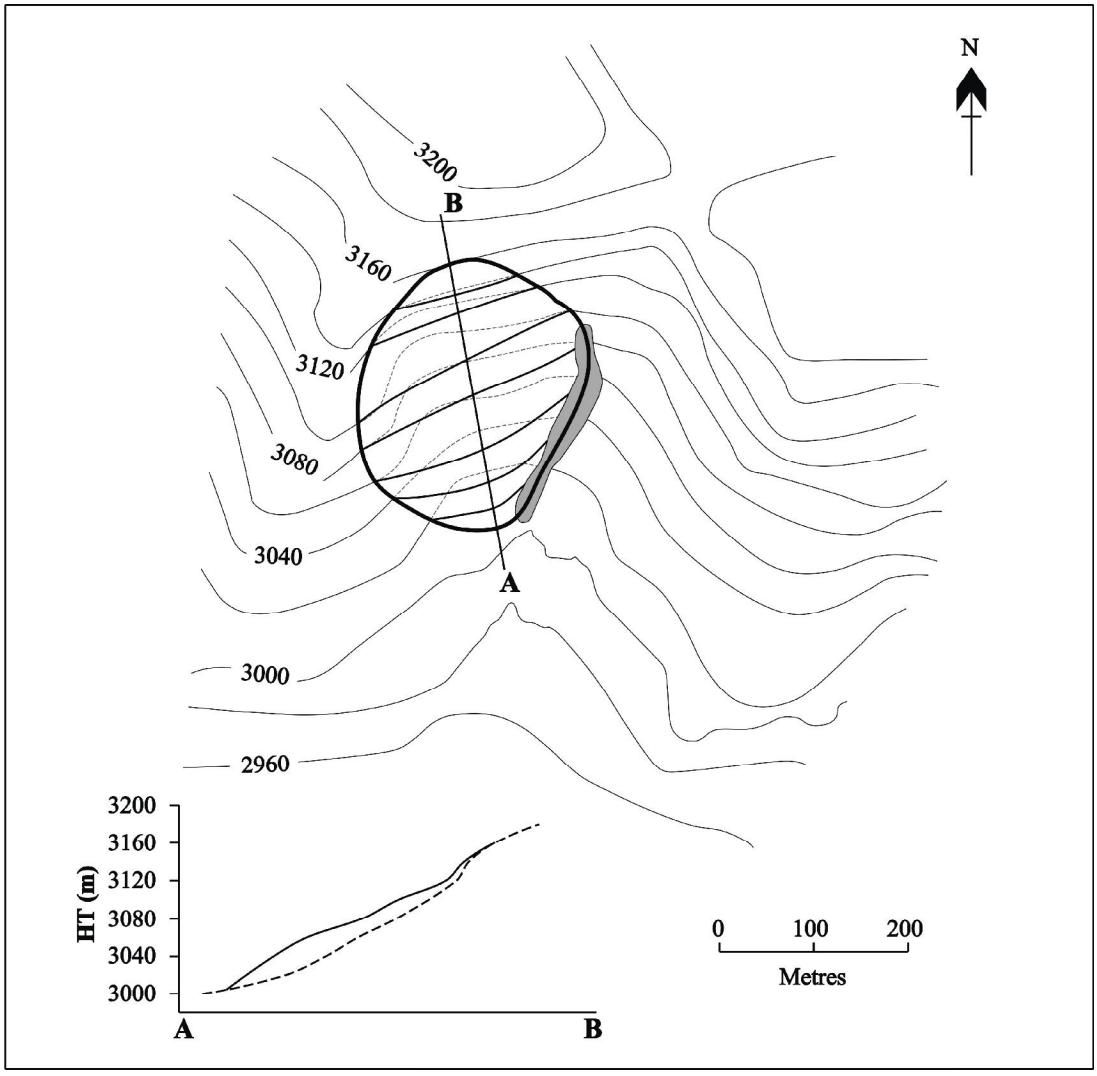


Figure 10.6 The reconstructed three-dimensional form of the possible palaeoglacier at Tsatsa-La-Mangaung.

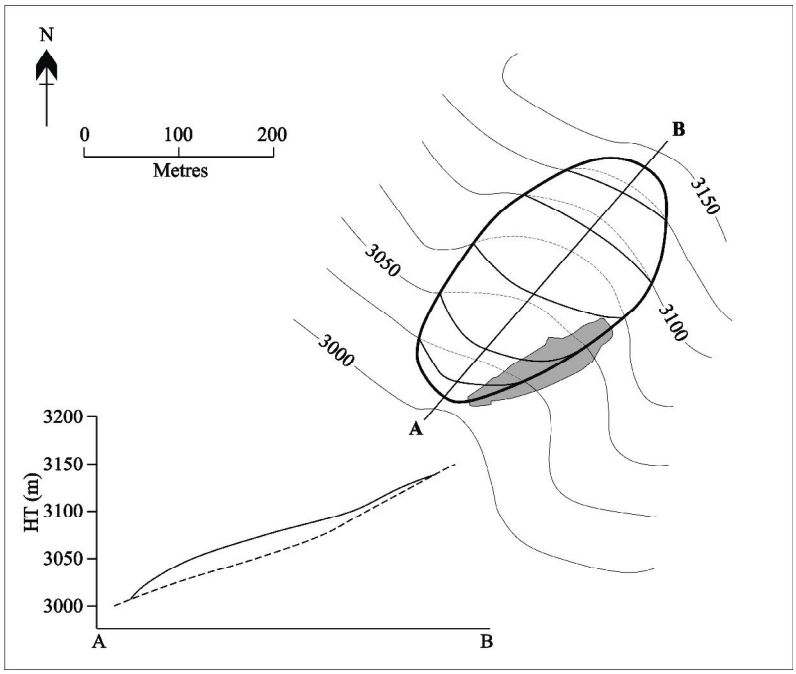


Figure 10.7 The reconstructed three-dimensional form of the possible palaeoglacier at Sekhokong Site 1.

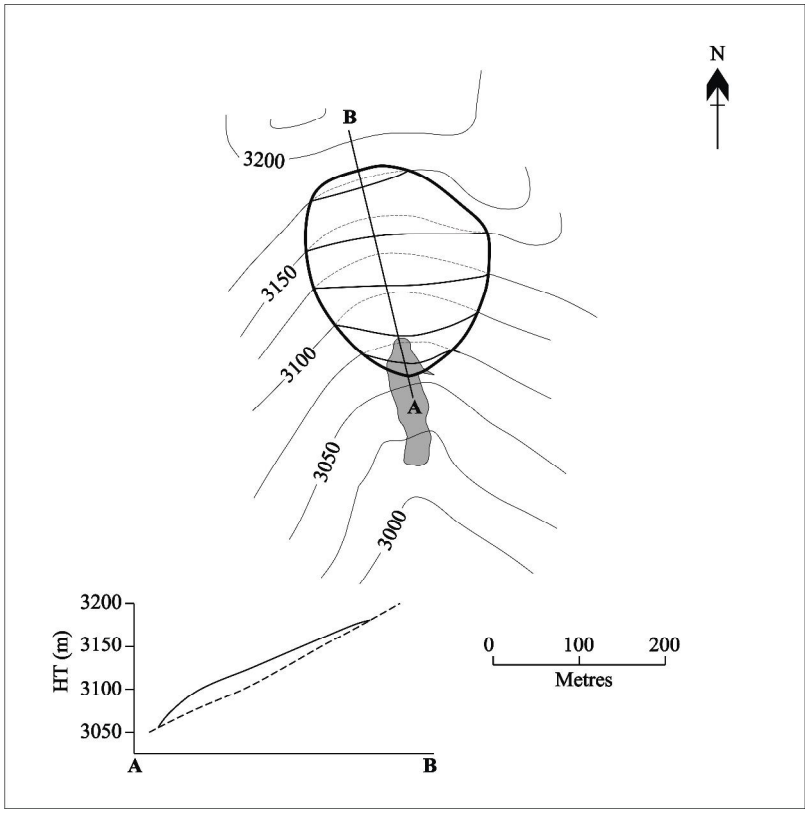


Figure 10.8 The reconstructed three-dimensional form of the possible palaeoglacier at Sekhokong Site 2.

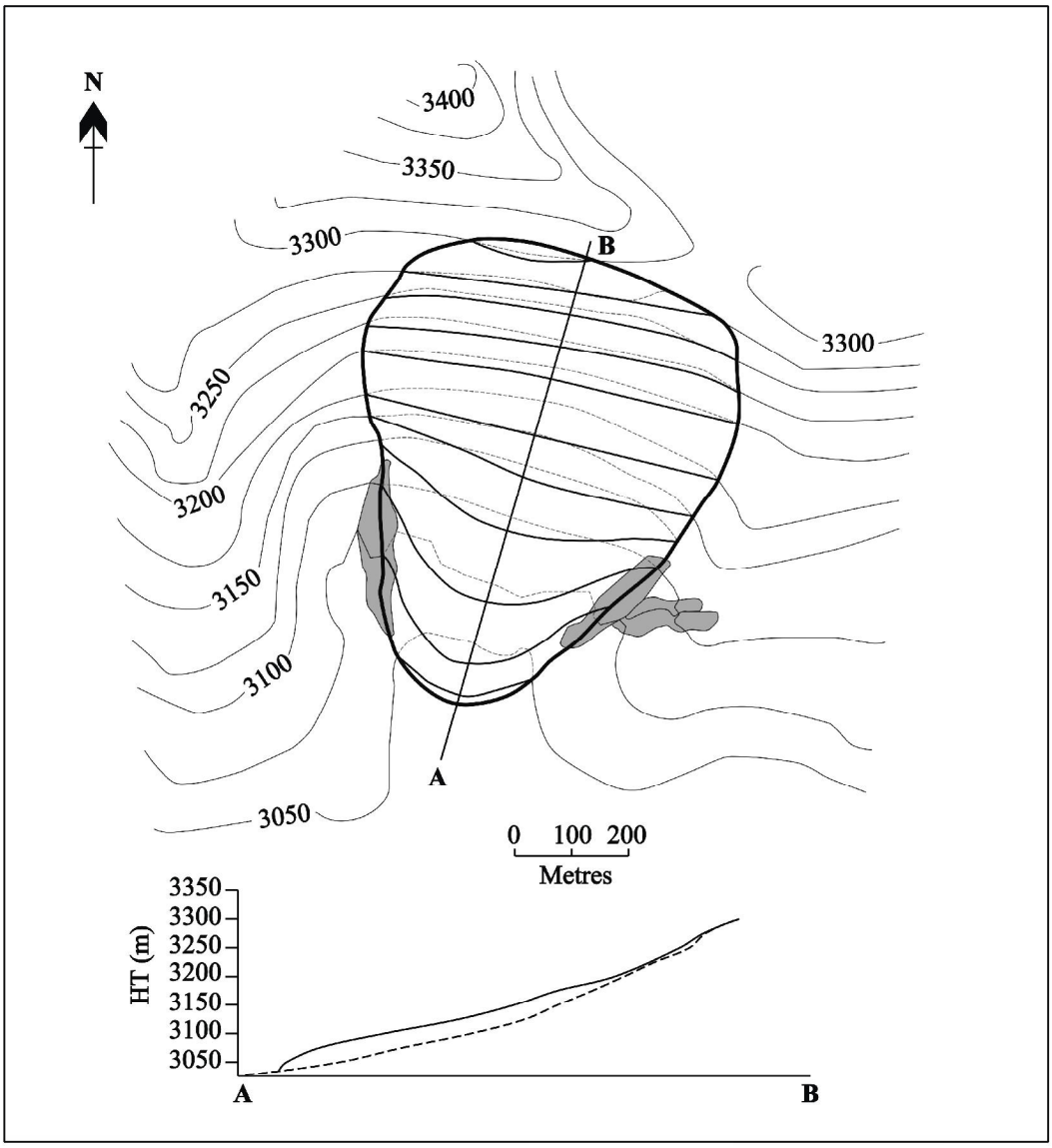


Figure 10.9 The reconstructed three-dimensional form of the possible palaeoglacier at Leqooa Valley.

Table 10.4 Ablation for each contour interval at Tsatsa-La-Mangaung, Sekhokong Sites 1 and 2 and Leqooa Valley.

Site	Contour interval	Ablation (mm)	Ablation (m)	Area (m^2)	Ablation (m^3)
Tsatsa-La-Mangaung	3070	73.5	0.074	12347	914
	3050	171.5	0.172	6703	1153
	3030	268.8	0.269	4102	1103
	3012.5	355.3	0.355	1957	695
Sekhokong Site 1	3062.5	210.65	0.211	11296	2383
	3037.5	333.15	0.333	5444	1813
	3015	455.65	0.456	1703	777
Sekhokong Site 2	3112.5	132	0.132	9557	1262
	3087.5	252	0.252	4192	1056
	3065	360	0.360	926	333
Leqooa Valley	3162.5	39.1	0.039	39735	1550
	3137.5	168.4	0.168	43109	7242
	3112.5	283.4	0.283	51707	14633
	3087.5	398.4	0.398	40970	16306
	3062.5	513.4	0.513	22682	11636
	3042.5	605.4	0.605	3068	1856

(2006) have suggested that the potential for ice deformation may exceed the predicted balance velocity of the glacier. Basal slip at Sekhokong Site 1 accounts for 33% of the total velocity, whilst at Leqooa Valley it accounts for 61%. In contrast, at Tsatsa-La-Mangaung and Sekhokong Site 2, ice deformation exceeds basal slip, indicating that in order to transfer mass through the equilibrium line, there would have been minimal need for basal slip (Carr and Coleman, 2006). All sites fall within the basal slip range suggested by Andrews (1975), Paterson (1981) and Carr and Coleman (2006). The palaeoglaciers at both Sekhokong sites are comparatively thin to the generally accepted minimum thickness of 30 m for glaciers (Gray, 1962; Shakesby and Matthews, 1993), however it is suggested that these glaciers may still have developed in this region, given that the transformation of snow to ice in temperate climates is rapid (Benn and Evans, 1998; Jania, 2004).

Table 10.5 Derived mass-balance characteristics at the four sites investigated.

	Tsatsa-La-Mangaung	Sekhokong Site1	Sekhokong Site 2	Leqooa Valley
ELA (m)	3085	3083	3140	3171
Temperature at ELA (ºC)	2.9	2.9	2.6	2.4
Accumulation at ELA	1570	1570	1500	1440
Ablation gradient (mm m^{-1})	4.9	4.9	4.8	4.6
Ablation (m^3 H_2O)	3865	4973	2651	52823
Mass flux (m^3 ice)	4210	5417	2888	57541
Cross-section at ELA (m^2)	3673	2144	2458	12855
Average total velocity at ELA (m a^{-1})	1.15	2.5	1.17	4.5
Basal sheer stress at ELA (bars)	0.6	0.50	0.58	0.64
Average deformation velocity at ELA (m a^{-1})	1.18	1.67	1.1	1.78
Basal slip as % of total velocity	0	33	0	61
Size of glacier (km^2)	0.052	0.042	0.039	0.39
Maximum glacier thickness (m)	31	22	22	35

The process by which snow is transformed to ice, and the time taken for the transformation to occurs, differs in warm temperate environments and in dry cold ones (Bennett and Glasser, 1996; Benn and Evans, 1998; Jania, 2004). Snow to ice transformation occurs at a depth of approximately 13 m below the surface, or within 3 to 5 years of burial, as a result of high accumulation and melt rates in a low altitude temperate area (Paterson, 1994; Bennett and Glasser, 1996; Benn and Evans, 1998). In warmer temperate environments where surface melting and percolation prevail, the transformation of snow to ice occurs rapidly (Lock, 1990; Menzies, 2002b; Jania, 2004). It is proposed that snow to ice transformation in the Drakensberg would have the potential to be rapid, given the large diurnal range in temperatures, which would accelerate melting and refreezing. It is therefore suggested that the minimum thickness of 30 m for glaciers may be an underestimation for the Drakensberg region, as is evidenced by ice formation occurring at 13 m at a temperate glacier (Paterson, 1994; Bennett and Glasser, 1996; Benn and Evans, 1998).

10.4 SNOWBLOW AND AVALANCHING

It is clear from Table 10.5 that the precipitation required for glaciers to be sustained exceeds the amount suggested in the region during the LGM (820 mm/annum). It is therefore necessary to investigate the effect that snowblow and avalanching may have had in providing the additional precipitation required.

It has been reported that avalanching is a major contributing factor for extra accumulation of snow onto a glacier, and in particular on influencing the mass balance of cirque glaciers (Dahl *et al.,* 1997). Zalikhanov (1975) reports that avalanches contribute between 10% and 65% of the accumulation on glaciers in the USSR, whilst Grosval'd and Kotlyakov (1969) and Tushinsky (1975) suggest that some glaciers are entirely dependant on avalanched snow. In addition, Inoue (1977) attributes 75% of the accumulation of a glacier in Nepal to avalanching. It has further been suggested that wind-blow and avalanching in the Caucasus is considerable, and that glaciers oriented transverse to prevailing winds received particularly significant accumulations from this source (Kotlyakov and Krenke, 1979).

Potential snow avalanching areas have been defined as those that slope at more than 20º, directly on the accumulation area of a former glacier, whilst potential snow-blow

areas have been defined as those slopes, within the catchment, that slope directly onto the accumulation area or a potential avalanching area (Sissons and Sutherland, 1976). It was also established that between 10% and 20% of accumulation on glaciers in the SE Grampians was from snow avalanching and snowblow (Sissons and Sutherland, 1976). Where there is a prevalent wind direction, snowblow may be critical in areas of marginal glaciation (Mitchell, 1996), and it has been shown that snowblow can be significant in determining glacier distribution and the ELA (Sissons, 1979, 1980a, 1980b; Sutherland, 1984; Leonard, 1989; Ballantyne, 1989b; Shakesby and Matthews, 1993; Ballantyne and Benn, 1994b). The small size of the glaciers in the Drakensberg (0.039 km^2 to 0.39 km^2) (Table 10.5) and the 88 m difference in ELA may indicate that the glaciers owe their existence to local factors.

In the Drakensberg, predominant winds are from the north and west in the winter (Tyson *et al.,* 1976; Freiman *et al.,* 1998; Sene *et al.,* 1998) and have been described in section 3.4.4. This is largely attributed to the high pressure system which prevails throughout the year over the interior of southern Africa (Tyson *et al.,* 1996). Work undertaken by Freiman *et al.* (1998) also describes dominant east-southeasterly winds which are attributed to a combination of regional plain-mountain winds blowing from the cooler plains towards the warmer escarpment, and local valley winds advancing upslope by anabatic flow during the day and in the early evening (Tyson *et al.*, 1976). Diurnal wind variation at Sani Pass recorded by Preston-Whyte (1971) indicates a predominantly north-westerly wind during the night and a predominantly south-easterly wind during the day. Wind data obtained at Sani Top (provided by Prof. Stefan Grab for the year 2001) are presented in Figure 10.10. Monthly 12 hourly wind roses are used to determine wind direction and speed during the night time (1800 to 0600) and daytime (0600 to 1800). Night time winds throughout the year are predominantly westerly to north-westerly; whilst during the daytime they tend to be more southerly. This is similar to wind data from Sani Pass described by Preston-Whyte (1971), who also observed north-westerly winds between 2100 and 0900, whereas southeasterly winds were more predominant after 0900. During the winter months (May to August), there is a stronger westerly to north-westerly direction than during the summer months. Mean wind velocity indicates that August is the windiest month, with an average of 9.1 m s^{-1}. The wind is also slightly stronger during the daytime than at night time.

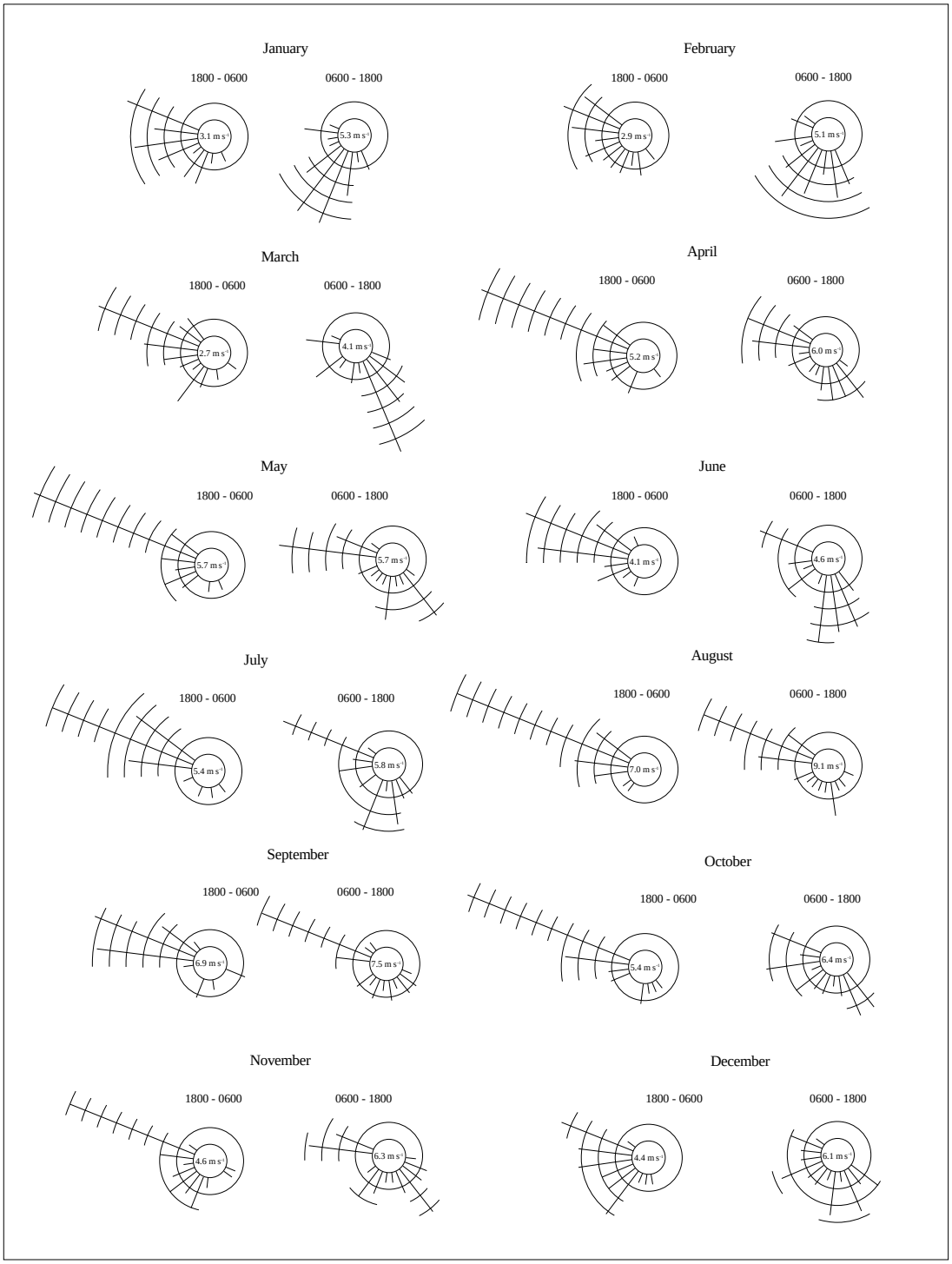


Figure 10.10 Surface wind roses for Sani Top. The arcs represent the 5% frequency intervals, as determined by dominant wind directions. Mean wind velocity is shown in the inner circle.

Calculation of the potential snowblow area for each palaeoglacier has followed the outline described in Sissons and Sutherland (1976), which considers all ground lying above the accumulation area of the glacier and sloping towards the glacier surface, as having the potential to contribute snow to the glacier surface. This approach has been used in all palaeoclimatic reconstructions of Loch Lomond Stadial glaciers in Great Britain (Sissons, 1980b; Sutherland, 1984; Ballantyne, 1989b; Mitchell, 1996) and by Dahl *et al.* (1997) in southern Norway (Figure 10.11). The snowblow areas for the four sites were digitized in order to calculate overall area and the significance of different wind directions for 15º sectors (Figure 10.12), which were drawn as polar plots (Figure 10.13). This information is further summarised in Table 10.6, which considers each 90º quadrant (N, E, S and W).

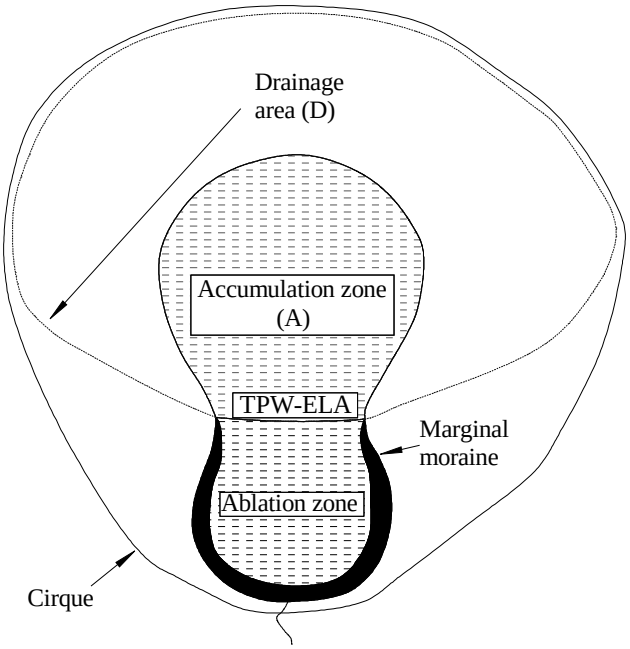


Figure 10.11 Schematic drawing showing the Drainage area (D) above the temperature-precipitation-wind equilibrium-line altitude (TPW-ELA) and the corresponding Accumulation zone (A) of a cirque glacier (after Dahl *et al.*, 1997).

It is apparent that none of the sites (with the exception of Sekhokong Site 1) receive any snow-blown precipitation from the south. This is because the glacier sites are on south-facing slopes and therefore the southern sector is below the equilibrium line. The Tsatsa-La-Mangaung and Sekhokong Site 1 glaciers would receive the majority of windblown snow from the north and east sectors, whilst the Sekhokong Site 2 and

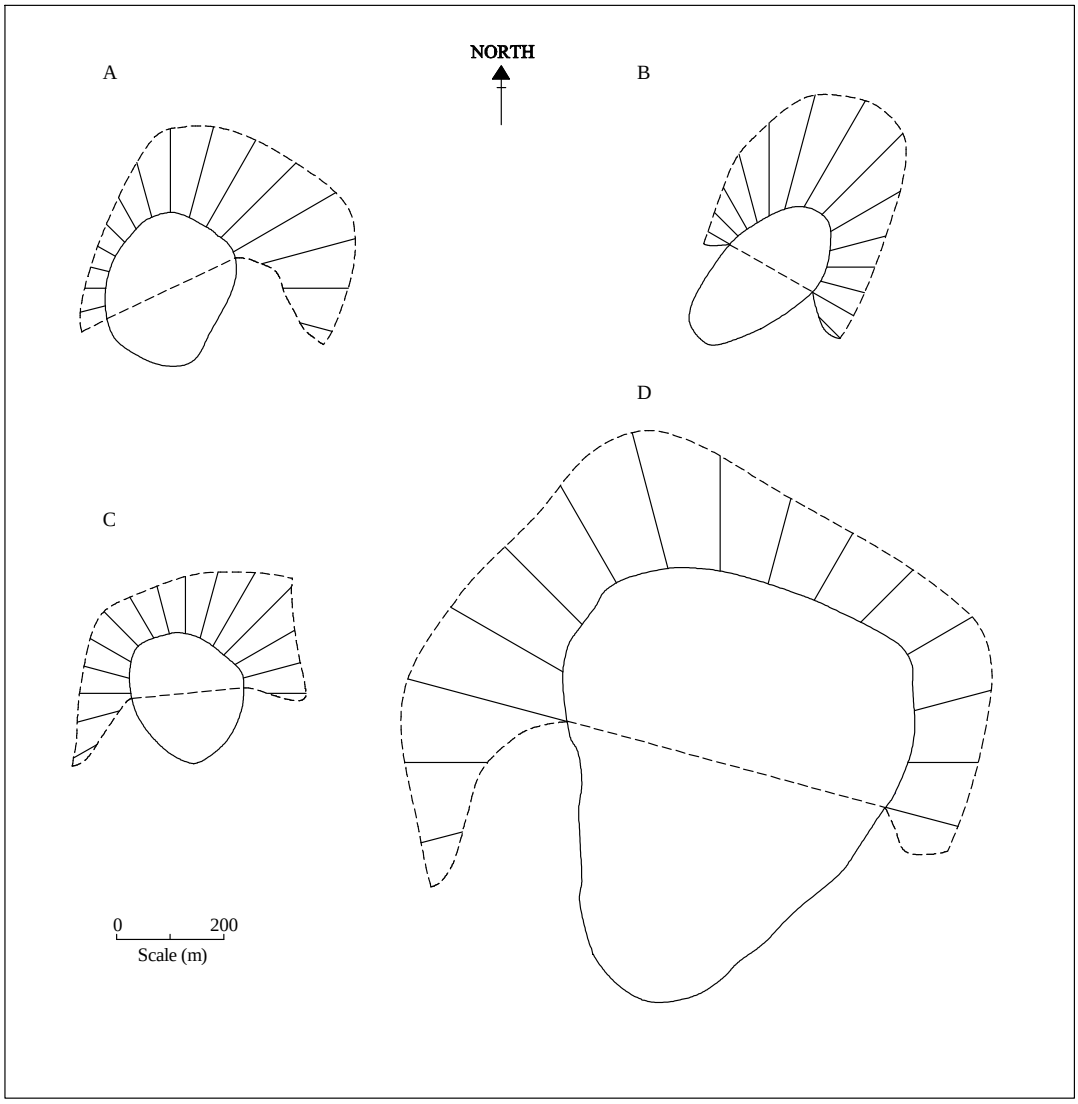


Figure 10.12 Potential snowblow areas for each former glacier (A = Tsatsa-La-Mangaung, B = Sekhokong Site 1, C = Sekhokong Site 2, D = Leqooa Valley).

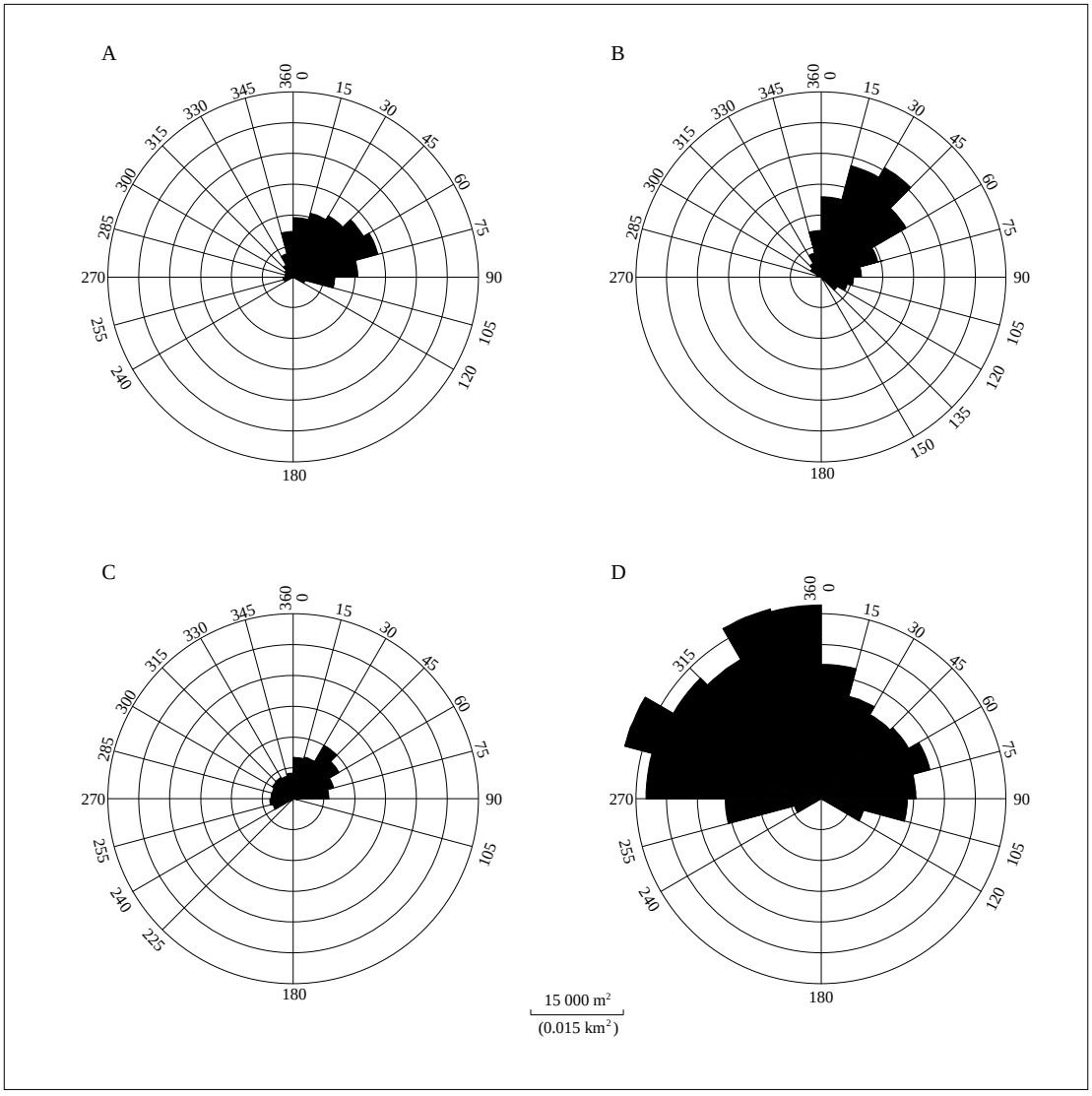


Figure 10.13 Polar plots of snowblow area and orientation for each palaeoglacier (A = Tsatsa-La-Mangaung, B = Sekhokong Site 1, C = Sekhokong Site 2, D = Leqooa Valley).

Leqooa Valley glaciers would receive windblown snow from the north, east and west sectors. The Leqooa Valley palaeoglacier also has the potential for the highest quantity of snowblow.

Dahl *et al.* (1997) suggest that there is a ratio between the drainage area above the ELA (D) and the reconstructed glacier-accumulation area (A) in the measurement of the potential for additional accumulation of snow by wind in cirques (Figure 10.11). A low D/A would indicate a minor additional contribution of windblown snow, whilst a high D/A ratio shows the potential for large amounts of windblown snow (Dahl *et al.*, 1997). The D/A ratio for the four sites ranges from 1.5 at Leqooa Valley to 4.7 at Sekhokong Site 1 (Table 10.7). This ratio may then be used to indicate winter precipitation inclusive of windblown snow, which means precipitation during the LGM at the four sites would have ranged from 1230 mm/annum at Leqooa Valley to 3854 mm/annum at Sekhokong Site 1. The Tsatsa-La-Mangaung and the Sekhokong Site 1 palaeo-glaciers appear to owe their existence to north easterly winds, where between 92% and 96% of the potential snowblow area is constrained within the north and eastern sectors. However the palaeo-glacier at Sekhokong Site 2 and Leqooa Valley would have also been influenced by westerly winds, with potential snowblow areas within the north, east and western quadrants, suggesting a dominant wind direction from the west-northwest. Contemporary snow-bearing winds are from the southeast and southwest (Nicol, 1973; Grab, pers Comm., 2006). The snowblow reconstructions however indicate that the snow-bearing winds may have had a more northerly direction during the LGM.

As discussed previously, snow avalanching may be an important source of accumulation on glaciers (Grosval'd and Kotlyakov, 1969; Tushinsky, 1975; Zalikhanov, 1975; Dahl *et al.*, 1997). Avalanches are initiated by storms, particularly heavy snowfall, and may occur through direct action (only involving new snow) or climax action (instabilities which develop within the snowpack over several days) (Owens, 2004). Snow avalanches tend to occur on the lee side of slopes on angles ranging between 30º to 45º (Owens, 2004). However, Sissons and Sutherland (1976) suggest that potential snow avalanching areas may occur on slopes exceeding 20º. The potential for snow avalanching is greatest at the Tsatsa-La-Mangaung site as a result

Table 10.6 Values of potential snowblow for each 90º sector.

Glacier	Area (km^2)	ELA (m)	Total snowblow area (km^2)	Snowblow area by 90º sectors expressed as percentage of total area N (316º - 45º)	E (46º - 135º)	S (136º - 225º)	W (226º - 315º)
Tsatsa-La-Mangaung	0.052	3085	0.098	46	46	0	8
Sekhokong Site 1	0.042	3083	0.085	60	36	0.4	3.6
Sekhokong Site 2	0.039	3140	0.069	47	25	0	28
Leqooa Valley	0.39	3171	0.32	45	22	0	33

Table 10.7 D/A ratios for the four sites.

	Tsatsa-La-Mangaung	Sekhokong Site 1	Sekhokong Site 2	Leqooa Valley
Drainage area (km^2) (D)	0.098	0.085	0.069	0.32
Accumulation area (km^2) (A)	0.026	0.018	0.019	0.21
D/A ratio	3.8:1	4.7:1	3.7:1	1.5:1
Precipitation with snowblow	3116	3854	3034	1230

of the steep slope (>45º), which falls within the drainage area, and the presence of the fault line which would channel avalanched snow onto the glacier (Figure 5.1 and 5.2).

Additional evidence for snow avalanching at this site is observed in the typically supraglacially derived debris in terms of the poorly consolidated particles and massive structure (Figures 5.4 to 5.6), the coarse, unimodal particle size distributions (Figures 5.12 to 5.14 and Table 5.1) and the predominantly angular clasts (Figure 5.22 and Table 5.9), whilst the other sites contain significantly greater proportions of fines and more rounded clasts. It has been suggested that in high mountain environments, snow avalanching provides the single most important source of supraglacial debris (Benn and Evans, 1998), which is represented in the sedimentological characteristics of the deposit at Tsatsa-La-Mangaung.

The glaciological approach to the interpretation of linear deposits at the Tsatsa-La-Mangaung, Sekhokong and Leqooa Valley sites provides strong supporting evidence for the occurrence of small glaciers during the LGM. According to Carr (2001), the analysis of reconstructed balance velocities of small glaciers and their likely components is the strongest indicator of a likely glacial/non-glacial origin, and all four palaeoglaciers are comparable with modern analogues.

Limited palaeo-precipitation has been a major concern with regards to glaciation in the Drakensberg (Hall, 2004); however necessary precipitation has been accounted for by including snowblow and avalanching. The exception to this is at the Leqooa Valley site, which has a relatively low snowblow ratio, indicating that windblown snow would have only made a minor contribution to the glacier. It has however, been suggested that the Leqooa Valley acts as a snow trap as a result of its topography, and that this area receives notable accumulations of snow compared to other areas of the Drakensberg (Mulder and Grab, 2002). In the eastern parts of southern Africa, the annual occurrence of cold fronts is approximately 43, and these predominantly occur towards late winter and early spring (Grab and Simpson, 2000). It has been suggested that during the Late Pleistocene, there was an increase in the frequency of cold fronts over the southern African coastal areas and adjacent interior (Van Zinderen Bakker, 1982). An increased frequency in cold fronts in winter and spring would have

produced heavier snowfalls and in turn increased the snow cover at high altitudes (Grab and Simpson, 2000).

The fact that the Drakensberg falls within the summer rainfall region is still somewhat problematic, given that snowfalls do not contribute the majority of total precipitation. The glacier reconstruction modelling has shown that given a 6ºC drop in temperature during the LGM, glaciers could have existed, however this is based on greater levels of precipitation than previously estimated. It has been suggested that a Southern Oscillation pattern could explain the climate at the LGM in southern Africa, whereby the Southern Indian high pressure cell is displaced north-eastwards, initiating drier conditions in summer and wetter conditions in winter in the summer rainfall region of South Africa (Tyson, 1986). It is however, unlikely that an inversion of the seasonal rainfall would have occurred (Tyson and Preston-Whyte, 2000). It is suggested that precipitation (snowfall) was greatly increased during the LGM, more so than previously thought, and that the 30% drop in precipitation suggested by Partridge *et al.* (1999) and Tyson and Partridge (2000) may have been a major over-estimation for the Drakensberg region.

There are only two synoptic systems that can bring substantial moisture to the required altitudes in the high Drakensberg. These are cut-off lows and very deep ridging highs (Washington, pers comm., 2006). Cut-off lows are an intense form of westerly trough, which deepen into a closed circulation and extend downwards to the surface, becoming displaced equator-wards (Tyson and Preston-Whyte, 2000; Dube, 2002). They are unstable systems that move westwards with increasing height (Tyson and Preston-Whyte, 2000). These cut-off lows contribute a significant amount of early- and late-season rainfall in South Africa (Taljaard, 1985, 1986), where peak times are in March to May and September to November (Tyson and Preston-Whyte, 2000). The combination of a cut-off low and a ridging anticyclone to the south produces widespread rainfall over South Africa (Dube, 2002). Ridging highs are associated with a westerly wave at 500 hPa and bring about rainfall along coastal margins (Vogel, 2000; Tyson and Preston-Whyte, 2000). The orographic effects of the Drakensberg result in high rainfall in these areas (Vogel, 2000). Light orographic rain can occur at any time on the north-east and south-east slopes of the Drakensberg Mountains as a result of warm maritime air approaching from the Indian Ocean (Van Zinderen Bakker and Werger, 1974; Sene *et al.*, 1998). Rainfall for the Drakensberg

region may be severely underestimated given the lack of rainfall measuring stations on these slopes. Ridging highs account for 16% to 25% of the rainfall variance over South Africa and produce extensive cloud cover along the southern and eastern coast and inland areas, which may rise above the Drakensberg Mountains and bring rainfall from October to May (Tyson and Preston-Whyte, 2000). An increase in these weather systems may have brought about sufficient winter precipitation in order to sustain the survival of the palaeoglaciers.

10.5 SOLAR RADIATION

Although the glacier reconstructions have proved successful, it must still be established why moraine-like features only occur at select sites and not on adjacent slopes which reach even higher altitudes. In order to explain the occurrence of these features on south-facing slopes, as opposed to north-facing slopes, solar radiation was recorded from November 2002 to October 2003 at Nhlangeni, which is approximately 10 km north of Sani top, near the Sehonghong River (Figure 10.14 and Table 10.8).

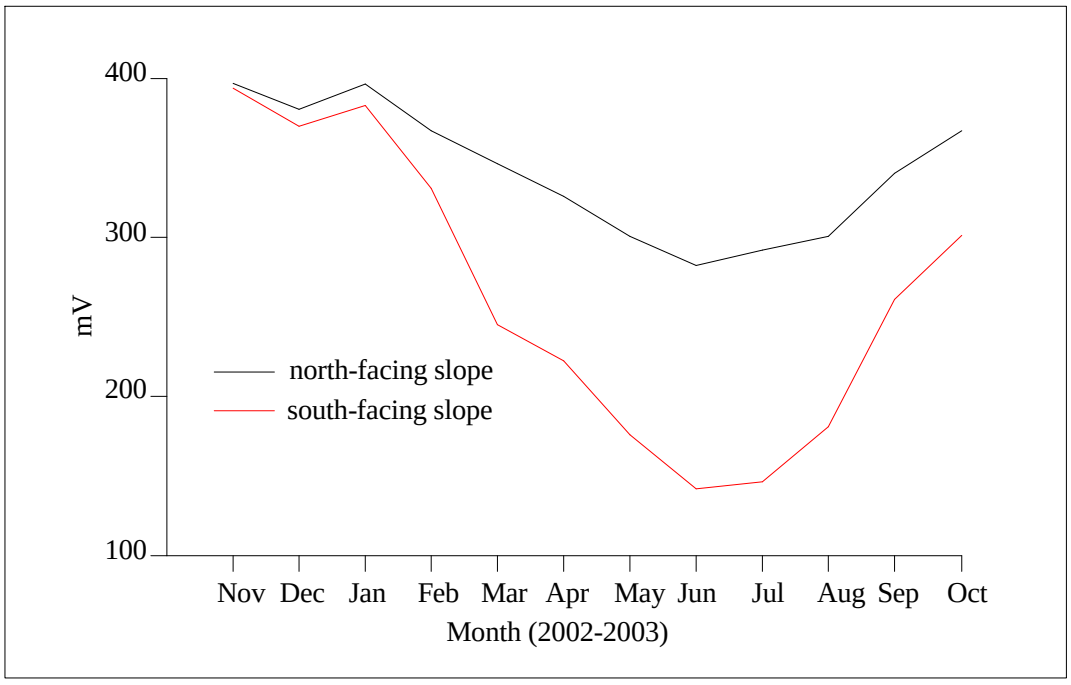


Figure 10.14 Mean monthly solar radiation on a north- and south-facing slope at Nhlangeni.

Figure 10.14 and Table 10.8 indicate that during the summer months, the amount of solar radiation received on both the north- and south-facing slope is very similar. In contrast, from March onwards, solar radiation values on the south-facing slope fall in proportion to those on the north-facing slope, to the point that in June and July, the south-facing slope receives 50% less than its northern counterpart. The higher solar radiation values in the summer months occur as a result of the longer days, whereas the lower winter values result from the longer passage of solar radiation through the atmosphere due to low solar altitudes (Tyson *et al.*, 1976). These solar radiation results represent an mV output, however radiation values in W m^{-2} for Lesotho have been presented (Ba and Nicholson, 2001). Values for fall between 225 W m^{-2} and 250 W m^{-2} in January, 175 W m^{-2} and 200 W m^{-2} in April, 125 W m^{-2} and 150 W m^{-2} in July and 200 W m^{-2} in October.

Table 10.8 Mean monthly solar radiation values for a north and south-facing slope at Nhlangeni.

Month	North-facing slope (mV)	South-facing slope (mV)
November (from 20th)	397	394
December	381	370
January	396	383
February	367	331
March	346	245
April	326	222
May	302	177
June	283	142
July	293	146
August	302	181
September	341	261
October (until 6th)	367	301

Additional solar radiation results are presented in Figure 10.18 and Tables 10.9 and 10.10 for various times of the day during the year. On the north-facing slope, solar radiation values are more or less constant between 9 am and noon, but fall rapidly by 3pm. This is more noticeable during the winter months where there is over a 50% drop by 3pm in July and August (Table 10.10). The results for the south-facing slope confirm results presented in Table 10.8, namely that these slopes receive >50% less insolation than their northern counterparts during winter, even at noon (Table 10.10). Although the solar radiation received on the north-facing slope is fairly constant by 9 am, it is slightly lower on the south-facing slope, reaching its maximum at noon. By

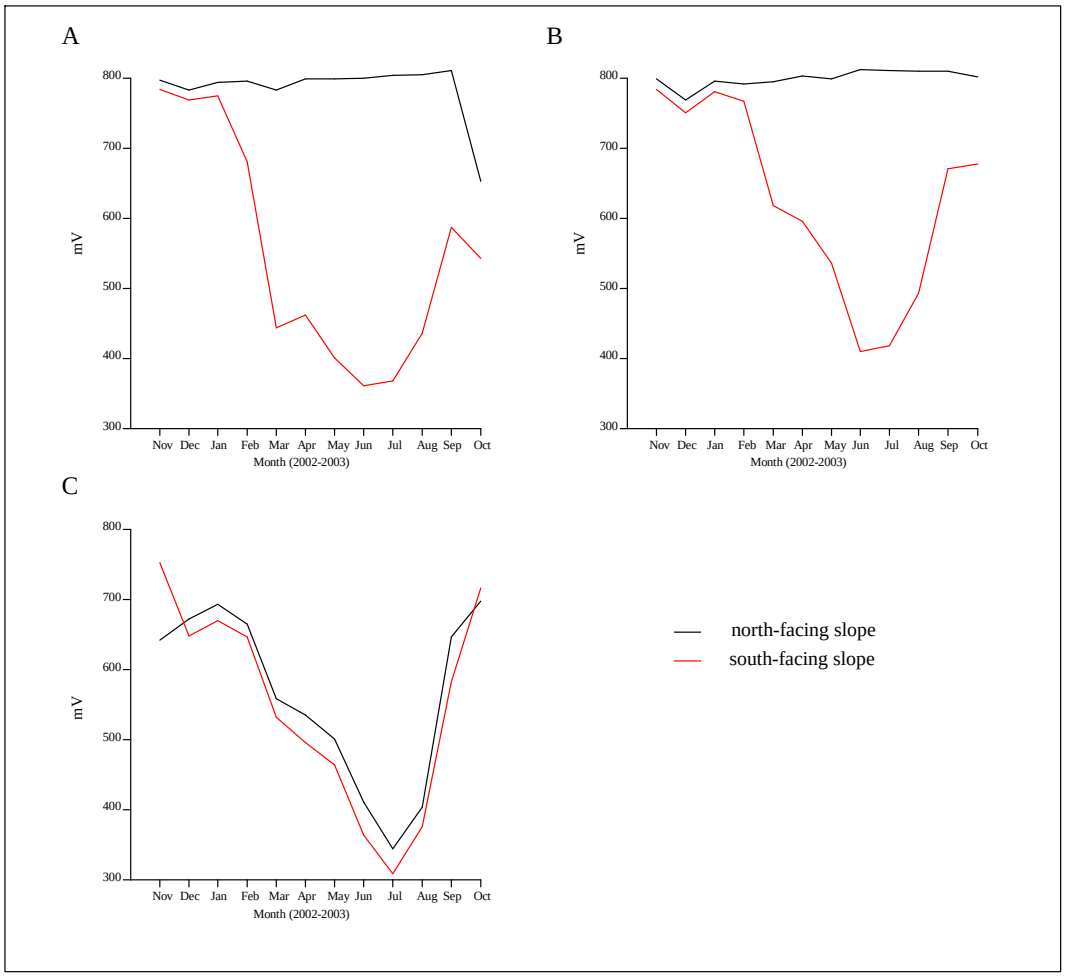


Figure 10.15 Solar radiation for the north and south-facing slope at Nhlangeni for various times of day (A = 9am, B = Noon, C = 3pm).

3pm both north and south-facing slopes receive considerably less solar radiation and they display almost identical trends.

The difference of 50% in summer and winter radiation values has also been expressed by Gopinathan (1991). Solar radiation was derived for Lesotho as a whole, based on data for one measuring station, whilst other values were estimated based on other parameters (Gopinathan, 1988a, 1988b and 1988c). Similar results to those obtained at Nhlangeni have been reported on a 10º north- and south-facing slope in the Drakensberg (Tyson *et al.*, 1976). They report that during the summer, the north-facing slope receives only slightly more radiation than the south-facing slope, whereas

Table 10.9 Solar radiation values for a north-facing slope at Nhlangeni at various times of the day.

Month	North-facing slope (mV) at 9 am	North-facing slope (mV) at noon	North-facing slope (mV) at 3 pm
November (from 20th)	797	799	642
December	783	769	672
January	794	796	693
February	796	792	665
March	783	795	559
April	799	803	535
May	799	799	501
June	800	812	411
July	804	811	344
August	805	810	404
September	811	810	647
October (until 6th)	653	802	698

Table 10.10 Solar radiation values for south-facing slope at Nhlangeni at various times of the day.

Month	South-facing slope (mV) at 9 am	South-facing slope (mV) at noon	South-facing slope (mV) at 3 pm
November (from 20th)	784	784	753
December	769	751	648
January	775	781	670
February	681	767	647
March	444	618	532
April	462	596	496
May	401	536	464
June	361	410	364
July	368	418	309
August	436	493	376
September	587	671	583
October (until 6th)	543	678	717

in winter, the south-facing slope receives about 50% less radiation, as observed at Nhlangeni (Figure 10.14 and Table 10.5). On slopes with a 30º angle, Tyson *et al.* (1976) report that a south-facing slope receives six times less radiation than its northern counterpart at midday and that as a result of the frequency of steep slopes in the Drakensberg, a complex spatial pattern of solar radiation may be expected, especially in winter when the sun is at low angles. This is further reinforced by Granger and Schulze (1977), who suggest that in winter, when solar altitudes are low, large variations in incoming solar radiation occur as a result of aspect and slope. In contrast, aspect is less important in the summer as a result of the high solar altitudes

and prevailing cloudy conditions, and therefore slope steepness is the major factor determining the amount of radiation received (Granger and Schulze, 1977). Furthermore, during the equinoxes, the influence of slope is more marked on the steeper southern aspects (Granger and Schulze, 1997).

The variation in annual solar radiation values are primarily due to changes in atmospheric conditions, such as cloudiness, water vapour and aerosol content (Ba and Nicholson 2001; Šúri and Hofierka, 2004). In Lesotho, the winter months are mostly cloudless, whilst the summer months are much cloudier, which is when most of the rainfall occurs (Tyson *et al.,* 1976; Gopinathan, 1991). This is because of the lower pressures, the shift in the ITCZ and the movement of moist air onto the continent from the east, which enables the development of thunderstorms (Vogel, 2000). The cloudy skies during summer reduce solar radiation levels in the eastern mountainous areas, whilst the western parts of Lesotho would usually receive higher radiation levels (Gopinathan, 1991). The palaeo-ablation season in Lesotho would have been from November to March, which is also the period when cloudy skies predominate (Gopinathan, 1991). This implies that solar radiation levels would be fairly low during this time as clouds strongly attenuate incoming radiation (Ba and Nicholson, 2001), hence less energy would be available for ablation (Benn and Evans, 1998). The influence of mountains upon shading of the sun at low altitude is typical of many locations within the Drakensberg (Tyson *et al.,* 1976). The summer months receive less sunshine than the winter months, as well as a rapid decline in the probability of sunshine after midday as a result of the frequent occurrence of afternoon cloud associated with orographic thunderstorms at this time of the year (Tyson *et al.,* 1976). The influence of afternoon clouds may account for the drop in solar radiation values in the summer months after noon. The drop in solar radiation received during the winter months after noon is likely to be as a result of shading from local topography.

Direct solar radiation is a function of latitude and time, and topography is the major factor modifying the local distribution of insolation, through variability in elevation, surface orientation and shadows cast by topographic features (Heggem *et al.,* 2001; Matzinger *et al.,* 2003; Chueca and Julián, 2004; Šúri and Hofierka, 2004). Orientation of the ground surface with respect to incoming solar radiation is particularly important in terms of favouring ablation on the local scale, and even more

in areas of marginal glaciation (Chueca and Julián, 2004). Small glaciers tend to be in delicate balance situations and are particularly impacted by aspect in their morphology and development (Chueca and Julián, 2004). Differences in the amount of solar radiation have been very important in the preservation or disappearance of small glacier remnants in the Pyrenees (Chueca and Julián, 2004), Kilimanjaro (Rosqvist, 1990), Mount Kenya (Young and Hastenrath, 1987), New Guinea and the South American Andes (Kaser, 1999). At Kilimanjaro, the radiation geometry causes a more extensive ice cover on the southern slopes, whilst the north side of Kibo receives greater insolation (Rosqvist, 1990; Young and Hastenrath, 1987). Glaciers reach low elevations on the western slopes, whilst the eastern slopes remain ice-free (Young and Hastenrath, 1987). They attribute this to the cloudiness in the afternoon and reduced insolation on the westward-facing slopes (Figure 10.16).



Figure 10.16 An example of afternoon cloud cover, covering Kibo (Kilimanjaro) on a south-western facing slope at 3950 m a.s.l. (photo by S. Mills).

Glaciers on Mount Kenya are also predominantly found on the western and southern slopes, as these are the most favourable areas for accumulation and survival of snow as a result of diurnal circulation and radiation geometry (Hastenrath, 1984; Young and Hastenrath, 1987). The highest summits of Mount Kenya are clear until late morning,

and then become progressively obscured by clouds that only dissipate in the late afternoon, which results in reduced insolation (Young and Hastenrath, 1987). This protects the glaciers and snow from the effects of ablation which are more pronounced on slopes which receive direct solar radiation (Young and Hastenrath, 1987). Work reporting on the importance of shade suggests that the geometry and mass balance of cirque glaciers are most likely to be influenced by local factors such as snow drifting and shading (Benn *et al.,* 2005). The influence of shade as well as the presence of afternoon cloud during the summer months may also have been the primary factors determining the presence of glaciers in certain parts of eastern Lesotho.

The variation in warm interglacial episodes and cold glacial episodes is driven by changes in the amount of solar radiation received from the sun (Bennett and Glasser, 1996). Periods with special climatic conditions have coincided with those of extreme solar activity, and those time periods with reduced solar activity have tended to be colder and synchronous with the expansion of the Earth's glaciers (Nesje and Dahl, 2000). It may therefore be expected that solar activity was reduced during the LGM, aiding the survival of the small palaeoglaciers in the Drakensberg.

Unfortunately, it was impossible to record solar radiation at the four sites. However, modelling hillshade can give an indication of the amount of insolation received at each site. Insolation may commence after sunrise, stop before sunset or cease temporarily if the slope is shaded for any given slope (Granger and Schulze, 1977). Hillshade based on shadows cast by topographic features was determined for the winter solstice, vernal and autumnal equinox and summer solstice, at various times of day (9am, noon and approximately 3pm) (Figures 10.17 to 10.20). Although these diagrams are only used to give an idea of hillshade, as lengths of shadows have not been calculated, they are nevertheless useful in explaining the likely occurrence of glaciers.

The most important time of year in terms of glacier survival is summer, as this is when ablation is likely to occur. Figures 10.17 and 10.18 indicate that the Tsatsa-La-Mangaung and Sekhokong Site 1 palaeoglaciers would have had almost half their area in shade at 9 am and again by 3pm during the summer solstice, which suggests that

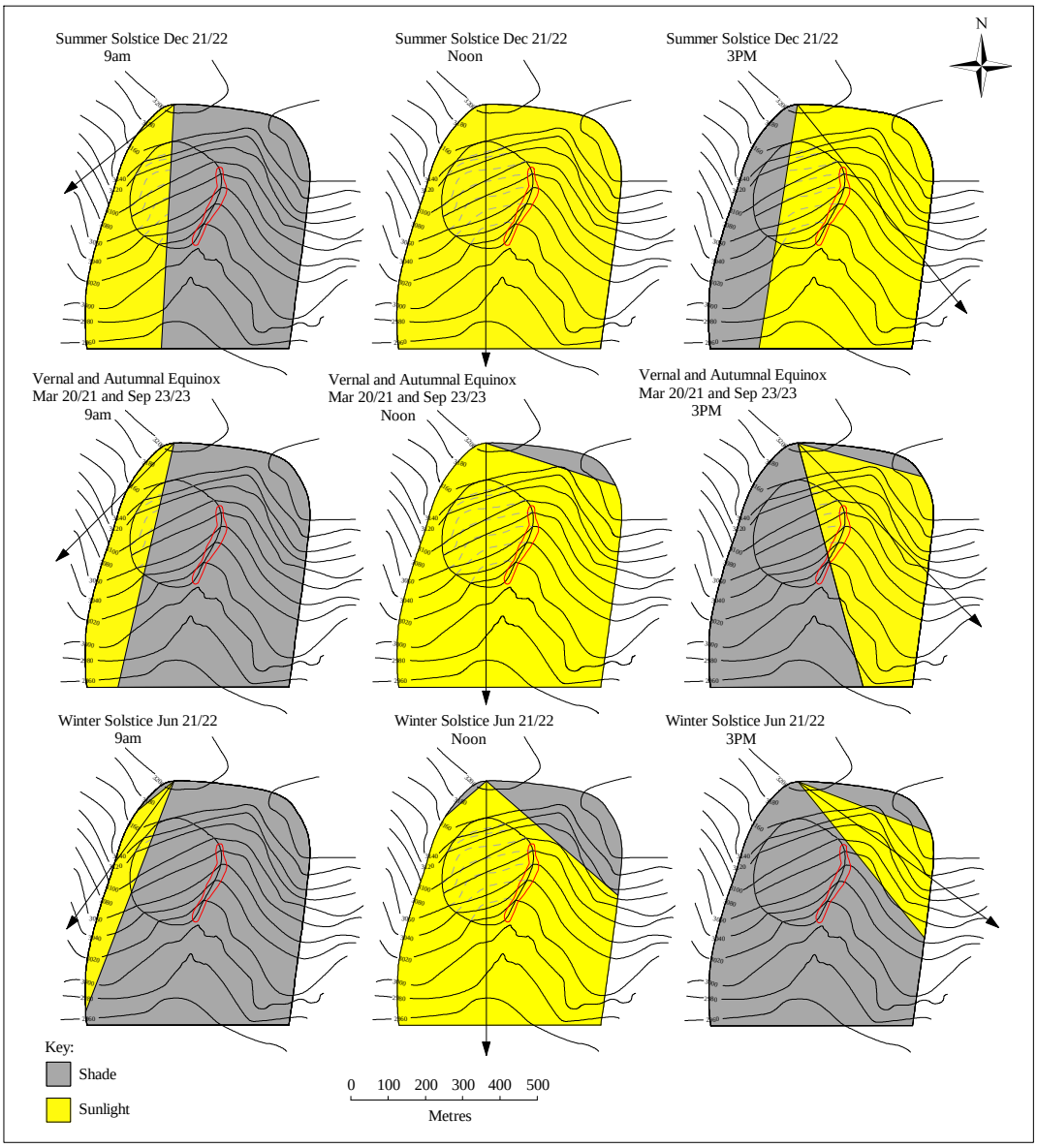


Figure 10.17 Map indicating areas of shade at different times of the day and year at Tsatsa-La-Mangaung. Note: Arrow indicates orientation of the sun.

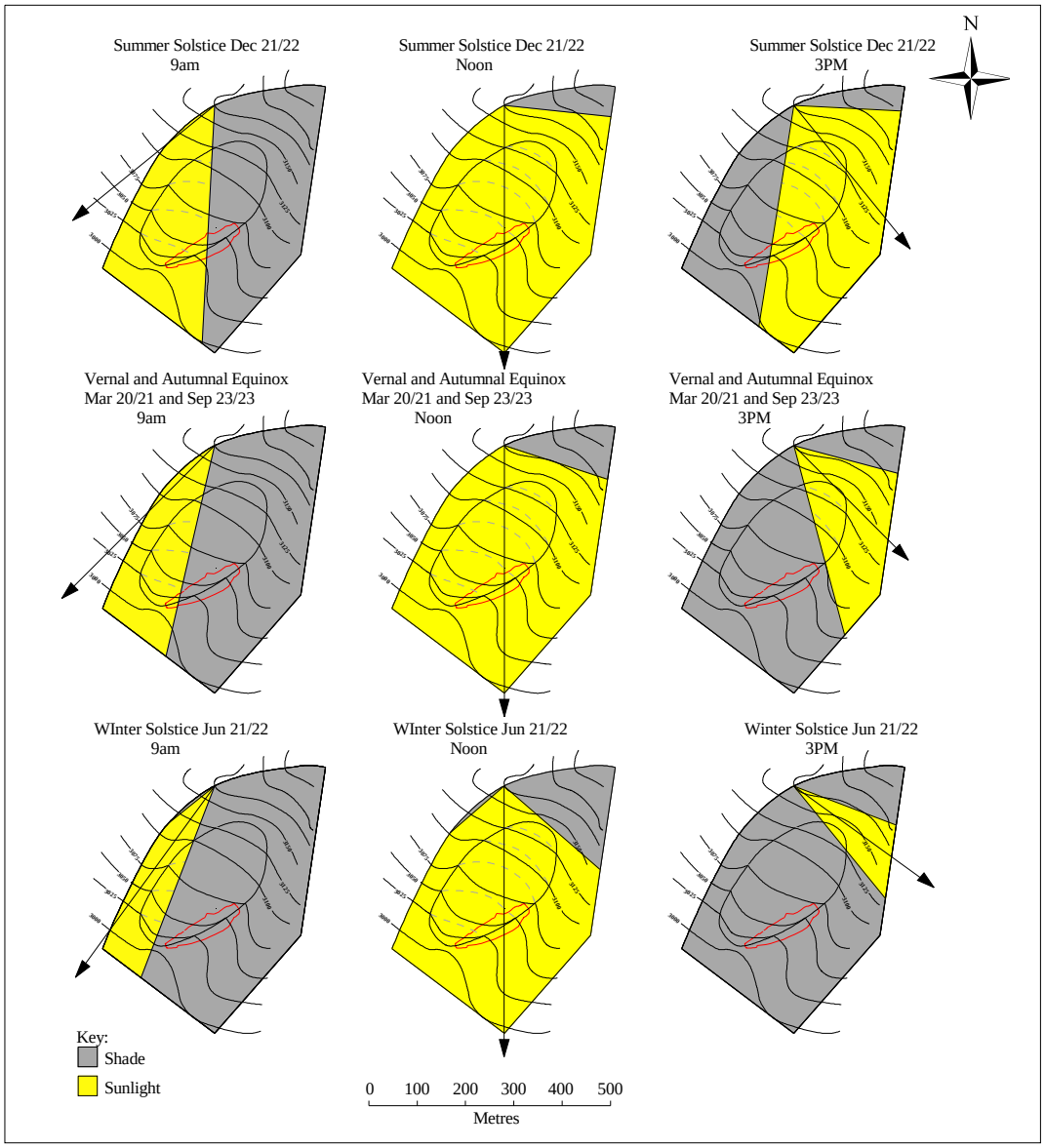


Figure 10.18 Map indicating areas of shade at different times of the day and year at Sekhokong Site 1. Note: Arrow indicates orientation of the sun.

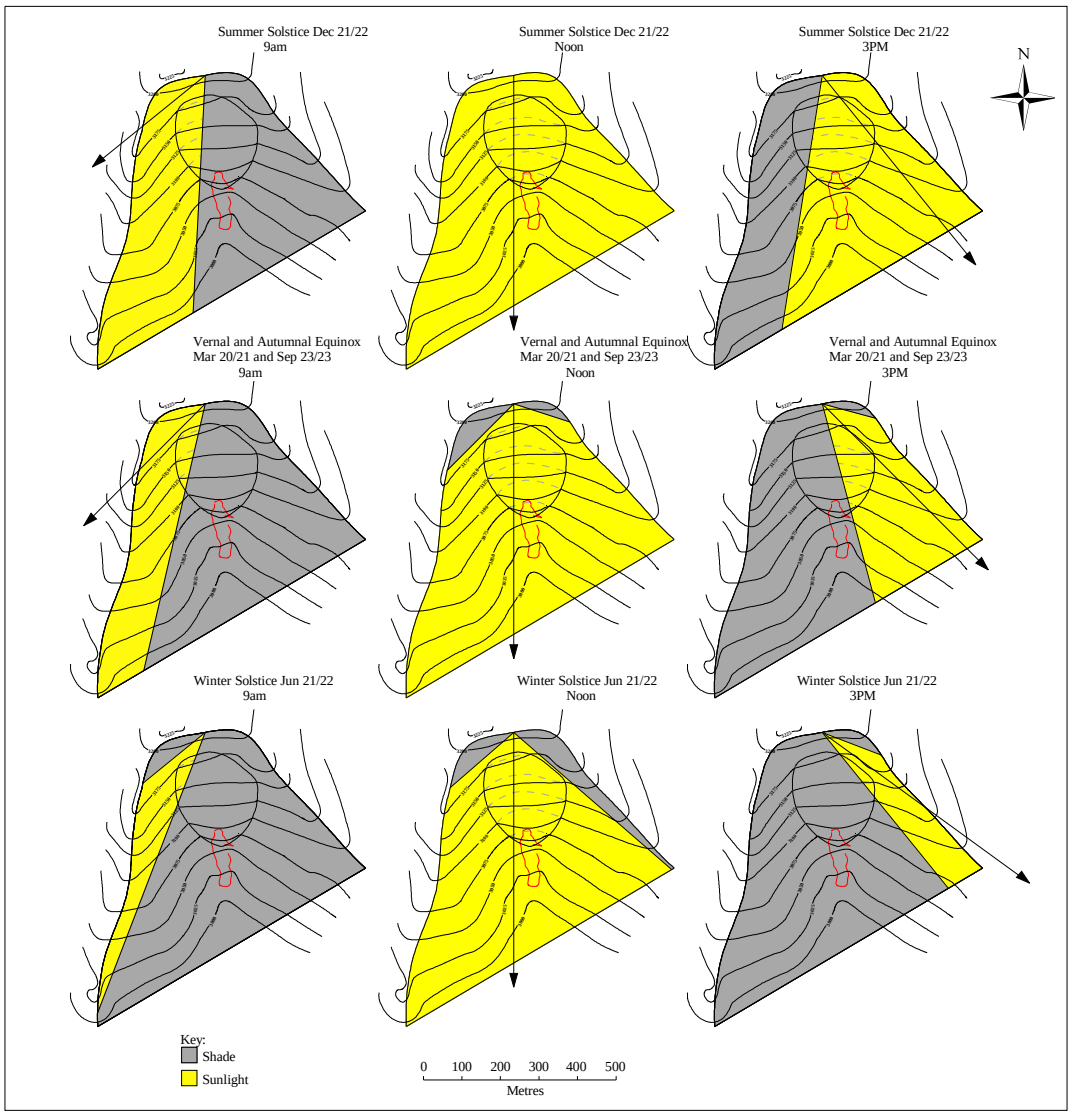


Figure 10.19 Map indicating areas of shade at different times of the day and year at Sekhokong Site 2. Note: Arrow indicates orientation of the sun.

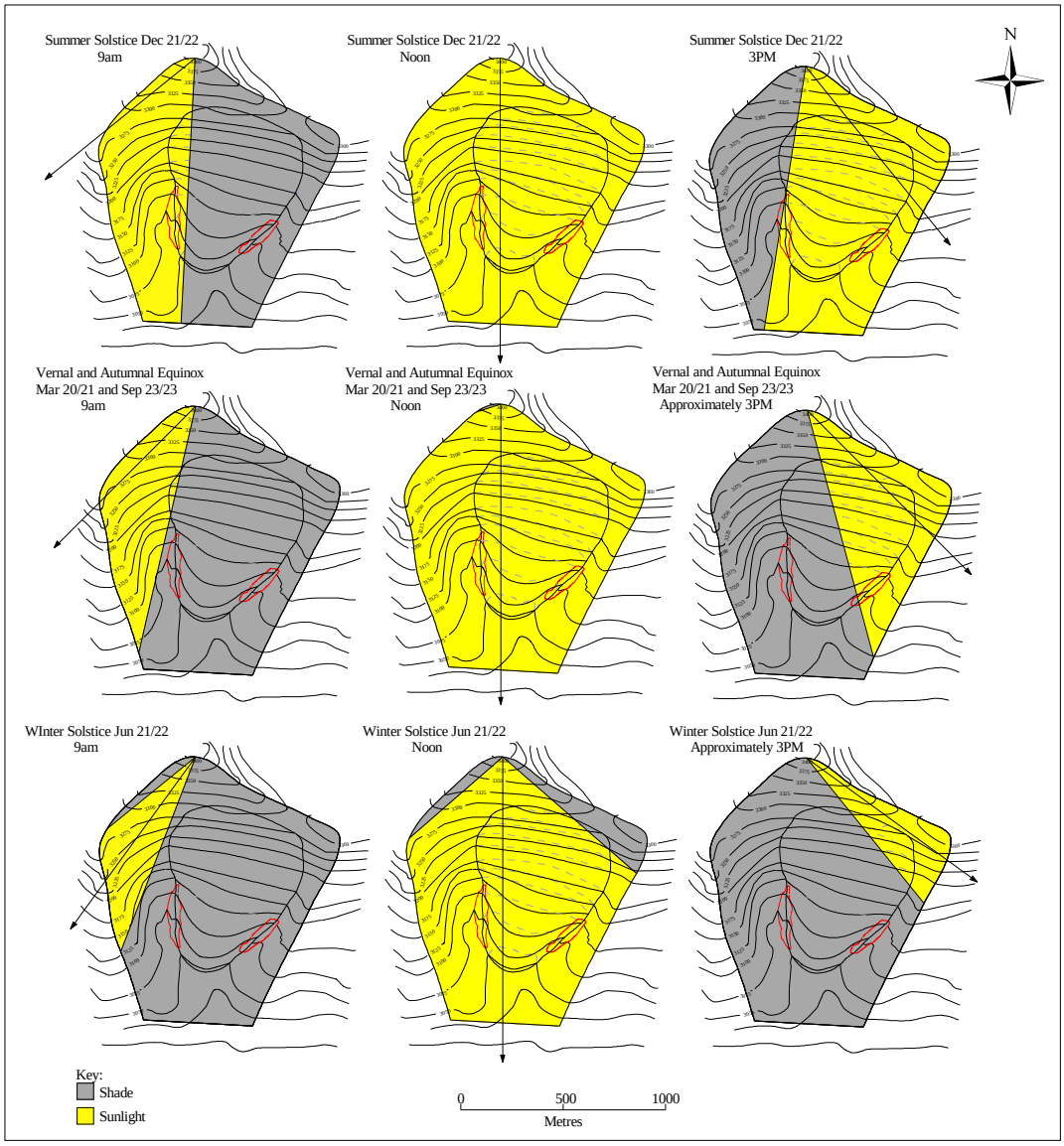


Figure 10.20 Map indicating areas of shade at different times of the day and year at Leqooa Valley.

Note: Arrow indicates orientation of the sun.

they would only be completely in sunlight for less than 6 hours of the day. This would greatly reduce potential incoming solar radiation received, and hence inhibit ablation. In contrast, the Sekhokong Site 2 and Leqooa Valley palaeo-glaciers would still have been almost entirely shaded by 9am, whilst it would have taken slightly longer for them to be completely shaded in the afternoon. They are none the less shaded in part by 3pm, suggesting that they would become fully shaded fairly rapidly as the angle of the sun falls (Figures 10.19 and 10.20). Afternoon clouds during the summer months would also aid in the reduction of solar radiation, as discussed previously. As would be expected, shading increases after noon, and around the vernal and autumnal equinoxes the Tsatsa-La-Mangaung and Sekhokong Site 1 palaeoglaciers would have approximately three quarters of their surface area in shade by 3pm, whereas the Sekhokong Site 2 and Leqooa Valley palaeoglaciers would have more than half their surface area in shade by this time. During the winter solstice, when the sun is at its lowest angle, the Tsatsa-La-Mangaung palaeoglacier would have still been almost completely shaded at 9am, and completely shaded by 3pm. The Leqooa Valley palaeoglacier would also have been in shade by 9am, and almost completely shaded by 3pm. Although the Sekhokong Sites are not entirely shaded at 9am and 3pm, they would only have a very small surface area in the sun. The shading from the morning and afternoon sun at all these sites may have been one of the determining factors in ensuring the survival of these glaciers.

The positions of the palaeoglaciers within these enclosed slopes would also have affected shading, as open valleys and north-facing slopes would have allowed for more incoming solar radiation and hence would have restricted the formation of glaciers. This is evident from Figure 10.21, which represents shading at the Sehonghong Site. This area is much more open and it lacks the same topographic restrictions as the linear deposit slopes. For this reason, the effect of the sun in relation to the mountain crest is altered for different times of the day. It is clear that throughout the summer solstice, this area is in direct sunlight throughout the day and only at approximately 6.30 pm would shading really become a factor. It is only during the winter that shading would have a significant effect at this site. However, shade is much more important during the ablation season, which may explain the lack of glacial evidence in this area as the rate of snow ablation would have been accelerated at this site.

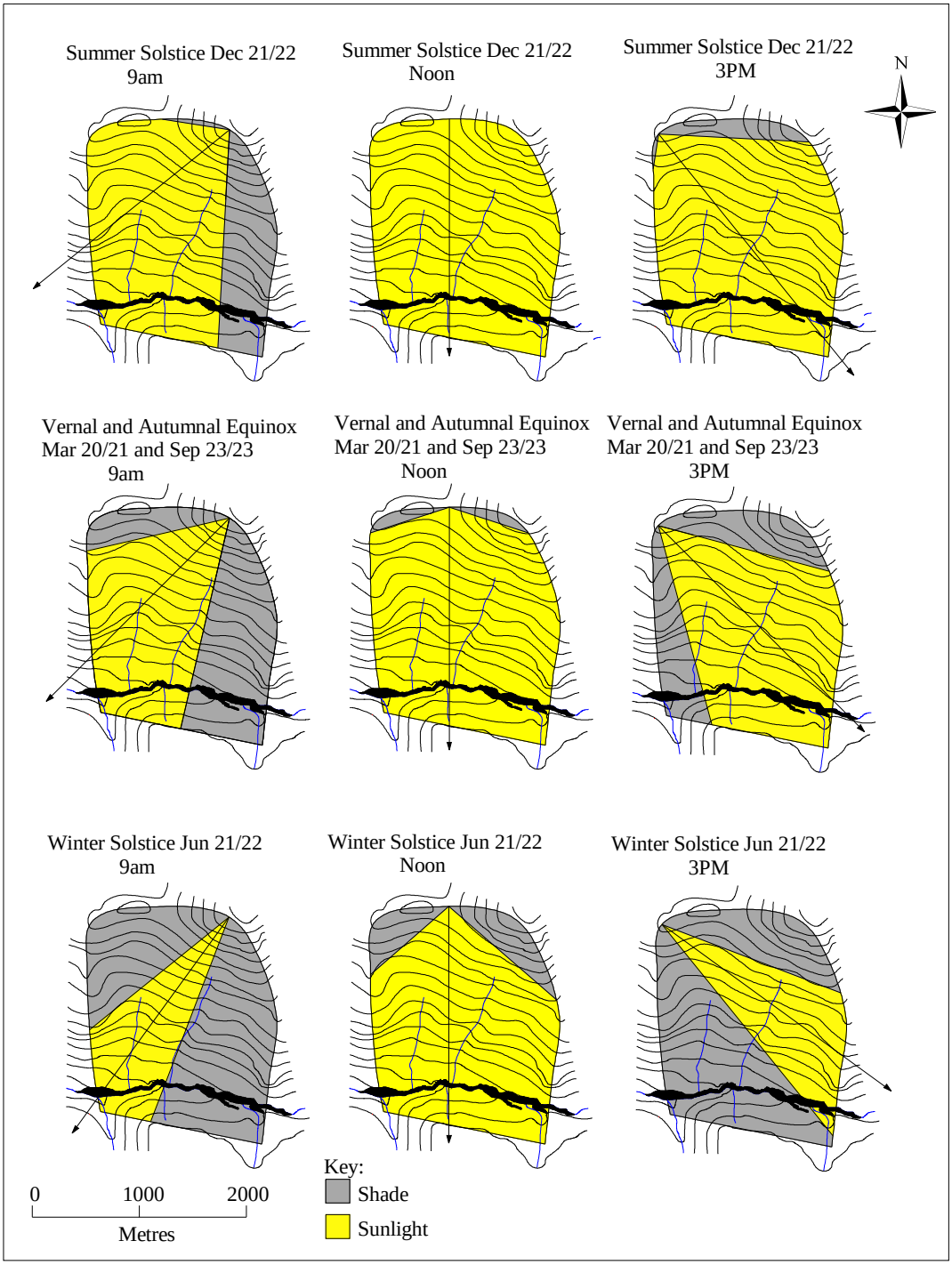


Figure 10.21 Map indicating areas of shade at different times of the day and year at Sehonghong.

Note: Arrow indicates orientation of the sun.

Although the same topographic controls from the east and west may be expected for north-facing slopes, meaning that they would experience some shade, the same cannot be said for the crestline above these slopes, as they are to the south and therefore would not create any shade. Diagrams are not shown for north-facing slopes, as these would be in direct sunlight throughout the majority of the day, given that sunlight would be reaching the slope directly. This is indicated in Figure 10.22, which compares the Leqooa Valley south-facing slope with its northern counterpart during the winter solstice. Both slopes are completely illuminated at noon, which would be expected, however even when the sun is at its lowest angle the north-facing slope is still completely in sunlight at 9 am and 3pm, whereas the south-facing slope along with the palaeoglacier are in the shade.

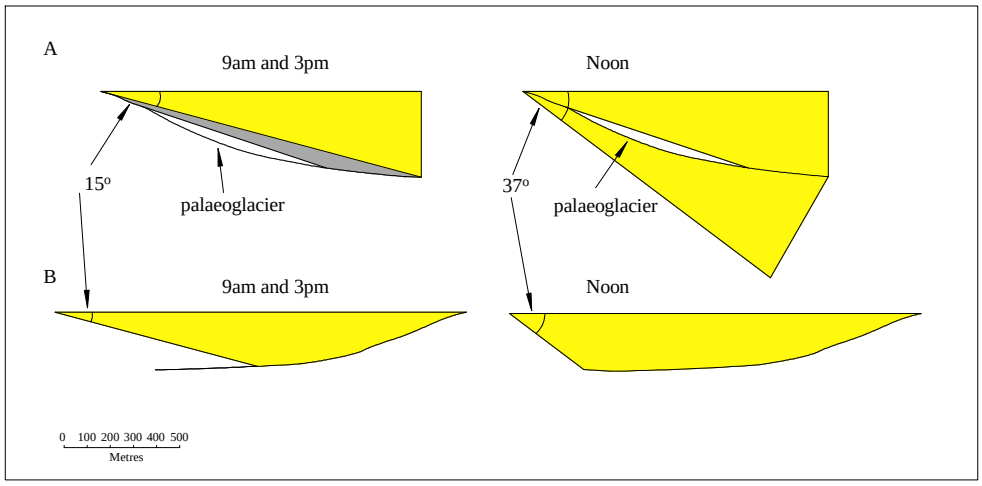


Figure 10.22 A comparison of direct sunlight reaching slopes at Leqooa Valley during the winter solstice (A = south-facing slope, B = north-facing slope).

10.6 CONTEMPORARY SNOW PATCH DISTRIBUTION

In order to further determine why only a few isolated areas were subject to glaciation, contemporary snow patch distribution must be taken into account. An indication of contemporary snow patch distribution is displayed in Figure 10.23 and 10.24. Figure 10.23 indicates snow accumulation at all sites and in particular substantial snow accumulation at Leqooa Valley and Sehonghong. Figure 10.24 shows remaining snow cover approximately 2 weeks later. It is clear that snow remains at all linear deposit sites; however there is a substantial reduction in snow cover at Sehonghong,

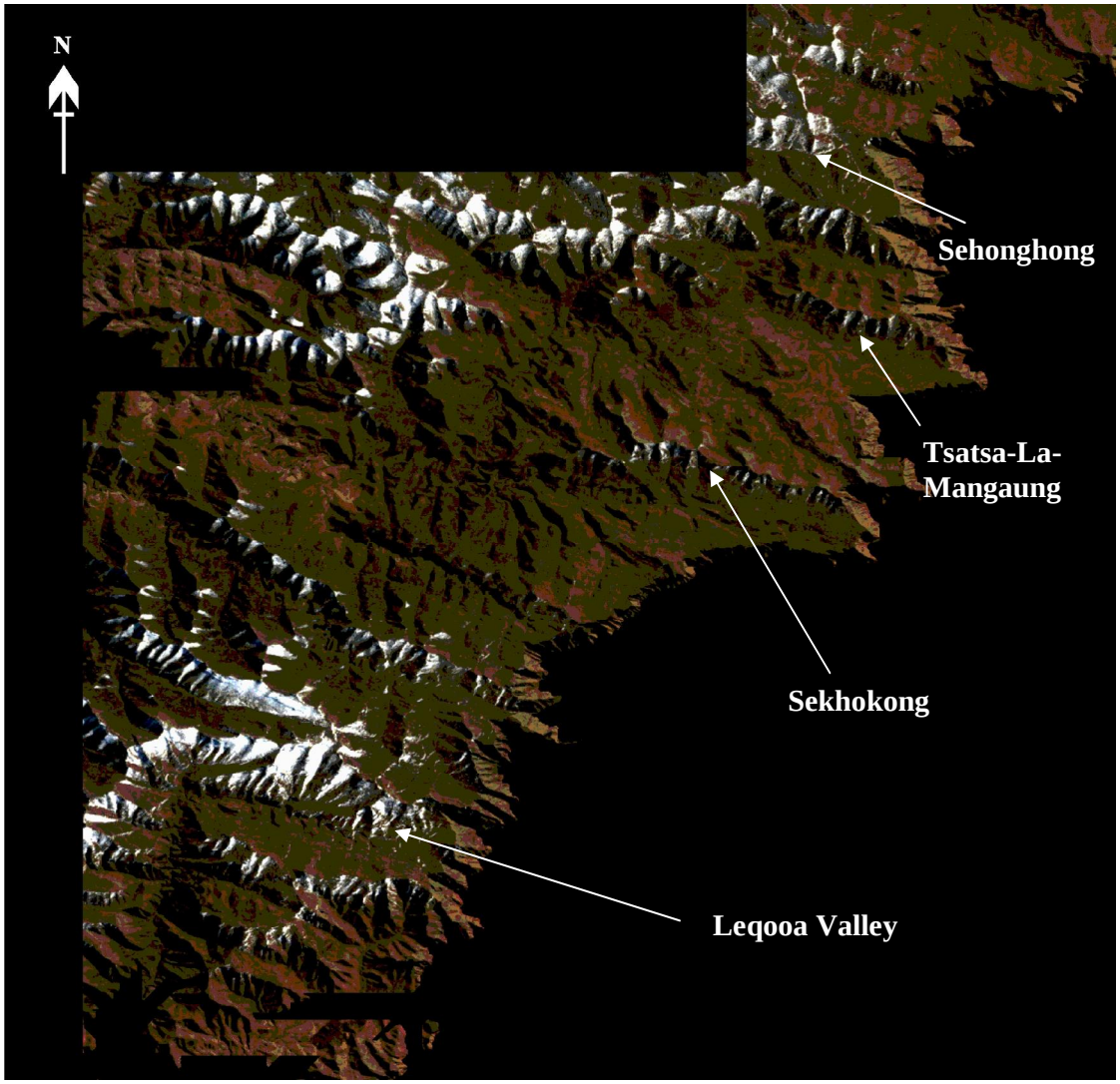


Figure 10.23 A real life colour composite image for the 3rd August 1990 (snow in white) (supplied by N. Mulder).

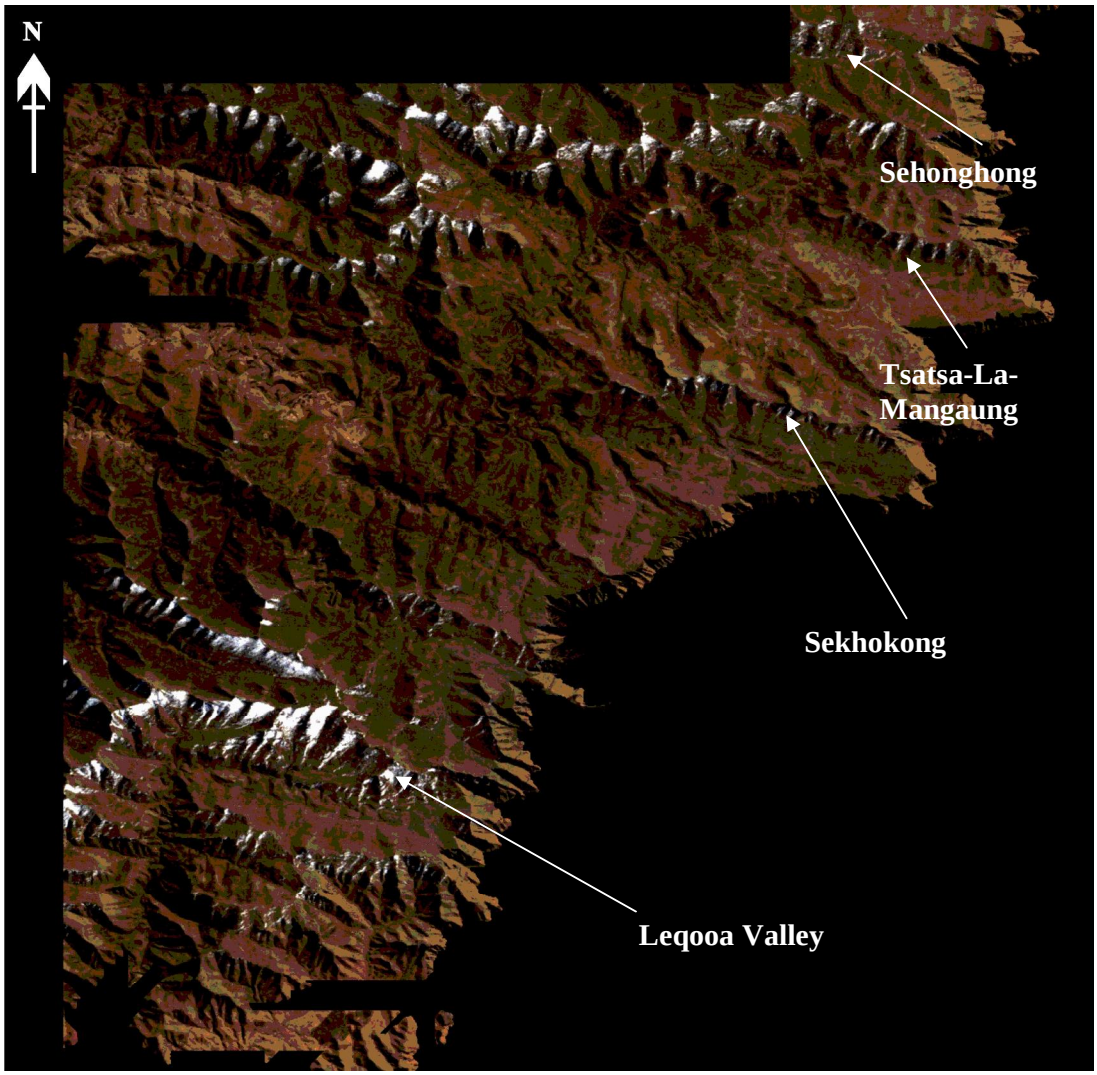


Figure 10.24 A real life colour composite image for the 19th August 1990 (snow in white) (supplied by N. Mulder).

which may be attributed to the greater insolation that this site receives, thus accelerating melting.

According to Mulder and Grab (2002), altitude does not appear to play a significant role in the amount of snow received, given that similar altitudes occur towards the west of the escarpment, whilst there are significantly fewer accumulations of snow in these areas (Mulder and Grab, 2002). Accumulations of snow occur westwards, as indicated in Figures 10.23 and 10.24; however this has been attributed to the shape of the Leqooa Valley acting as a snow trap and airflow up the Leqooa Valley. It is suggested that distance from the escarpment and aspect are predominant factors, as the amount of snow received diminishes with distance from the escarpment, and snow patches predominantly persist on south-facing slopes (Mulder and Grab, 2002). It is suggested that the lack of snow on north-facing slopes is due to the severe ablation of snow as a result of higher levels of insolation and the lack of wind-blown snow (Mulder and Grab, 2002). Areas to the north of Sani Top lack linear deposits such as those found at Tsatsa-La-Mangaung, Sekhokong and Leqooa Valley, and it is suggested that this is because these areas receive less snow (Mulder and Grab, 2002). Although areas north of the study sites do not contain such linear slope deposits, it may be that areas to the west of the Leqooa study sites also contain similar deposits, associated with the late-lying snow patches.

10.7 SUMMARY

The glacier reconstructions for the Tsatsa-La-Mangaung, Sekhokong and Leqooa Valley sites are comparable with modern analogues, suggesting that glaciers could have existed in these localities, given a 6°C temperature depression, as suggested by Heaton *et al.* (1986), Talma and Vogel (1992), Partridge *et al.* (1999), Tyson and Partridge (2000), Grab and Simpson (2000), Holmgren *et al.* (2003). Based on the analysis of the solar radiation results and contemporary snow patch distributions, it is clear that south-facing slopes receive considerably less solar radiation and more late-lying snow cover than their northern counterparts. It is suggested that topographic (e.g. altitude), microtopographic (e.g. slope gradient and aspect) and particularly microclimatic factors (e.g. late-lying snow cover, snowblow and solar radiation) play an important role in the distribution of the palaeoglaciers. It has been shown that areas below 3000 m a.s.l. are unlikely to have been conducive to glacier formation as a

result of a greater necessity for precipitation, whilst north-facing slopes do not favour the survival of late-lying snow patches owing to solar radiation being much more intense on these slopes. Adjacent areas to the study sites would be either too far from the escarpment to receive necessary snow, or lack the topographic shadows necessary to restrict incoming solar radiation, which hence increase snowmelt. It is therefore suggested that the High Drakensberg was within the glacial/periglacial equilibrium zone during the LGM; however the glacial zone was restricted to very few sites which comprised a delicate balance between necessary topographic and climatic factors to ensure the survival of small glaciers (Figure 10.25). It is possible that further slope deposits occur in the high Drakensberg, however it is suggested that these would occur principally on high-altitude south-facing slopes, where topographic features restrict insolation, and where there is the greatest amount of late-lying snow.

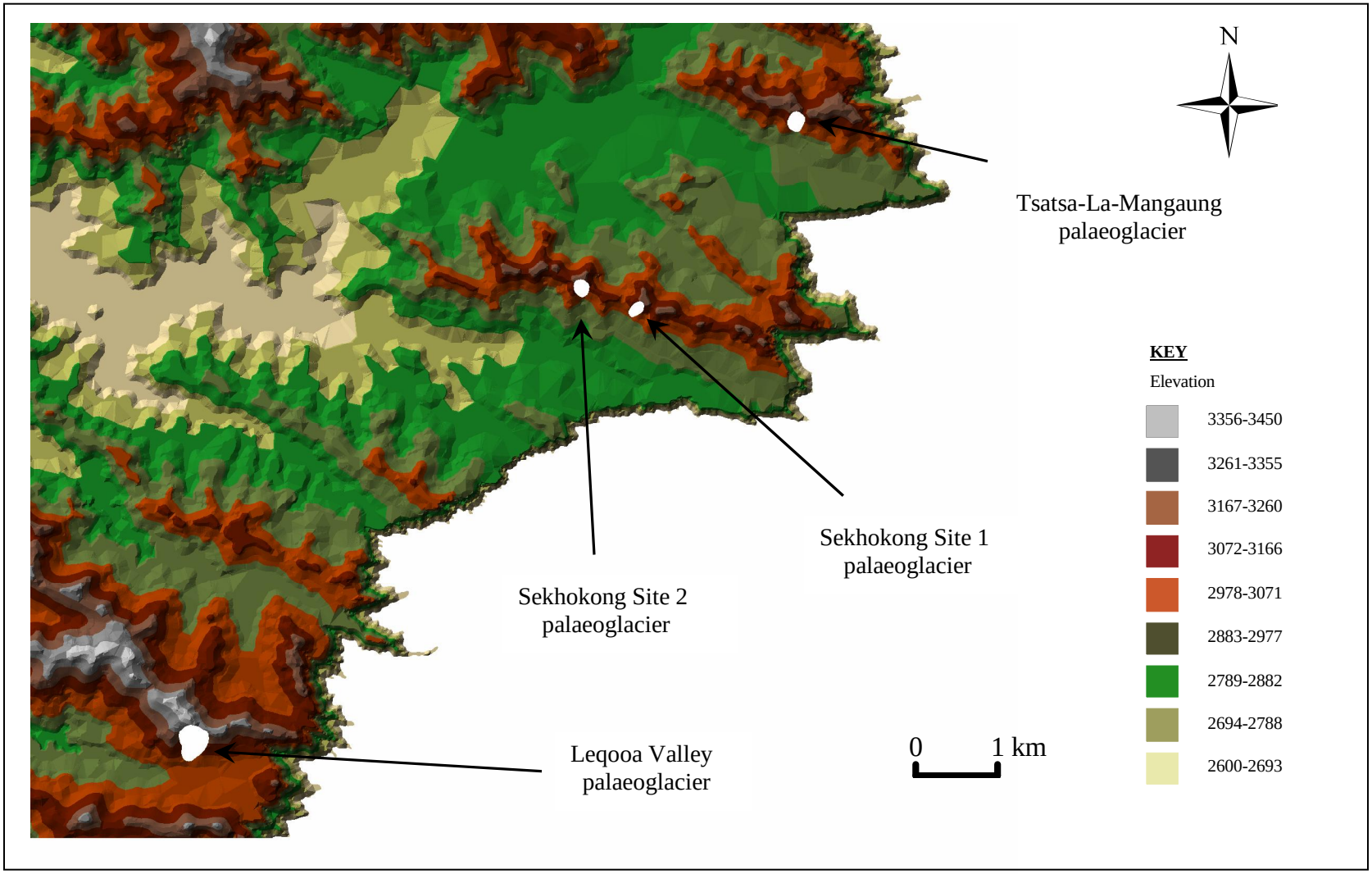


Figure 10.25 DEM of the position of the palaeoglaciers in relation to each other.

CHAPTER 11

Conclusion

11.1 INTRODUCTION

The objective of this research project was to ascertain the likely geomorphic processes that produced the debris deposits located at a few isolated localities in the high Drakensberg. The aims were to identify and compare the morphology of the various slope deposits found in eastern Lesotho, analyse the sedimentology and micromorphology of the deposits and in so doing, provide some understanding of the past geomorphic processes that operated at the sites, and finally to undertake a multi-faceted methodological approach to evaluate the feasibility for past Quaternary glaciation in the high Drakensberg. The extent to which these aims and objectives of this research are realised will be discussed in this final chapter, along with the contribution this research will make towards understanding the Quaternary history of the Great Escarpment of southern Africa. Finally, this chapter will discuss the relevance of this research in an international context and identify directions for future research in this region.

11.2 EXTENT TO WHICH THE OBJECTIVE AND AIMS ARE REALISED

The likely geomorphic processes which produced the deposits at Tsatsa-La-Mangaung, Sekhokong Sites 1 and 2, Leqooa Valley and Sehonghong have been successfully determined through the use of morphological, sedimentological and micromorphological techniques. Various process origins were considered and rejected on the basis of whether the deposits suited the attributes typical to these various process mechanisms. AMS dating also enabled the time of deposition to be determined at the Tsatsa-La-Mangaung, Sekhokong and Leqooa Valley sites, and for the lower units of the Sehonghong Site 3 deposit. Furthermore, the process origins of the linear slope deposits were further quantified by undertaking a glacial reconstruction history for each site in order to determine whether glaciers could have existed in these areas. These sites were then analysed for solar radiation received and contemporary snow patch distribution, so as to resolve why only certain areas were conducive to glacier formation.

11.3 CONTRIBUTION TO THE QUATERNARY HISTORY OF THE GREAT ESCARPMENT

Contradicting views have long existed regarding the Quaternary glacial history of southern Africa, with several studies arguing for glaciation (e.g. Sparrow, 1967a; Marker and Whittington, 1971; Harper, 1969; Dyer and Marker, 1979; Borchert and Sänger, 1981; Sänger, 1988; Hanvey and Lewis, 1990; Marker, 1991; Lewis, 1996; Hall, 1994; Grab, 1996a; Lewis and Illgner, 2001; Mills and Grab, 2005), whilst others have presented alternative hypotheses for the occurrence of these features (e.g. Meadows, 1988; Sumner, 1995; Grab and Hall, 1996; Grab, 2000a; Boelhouwers and Meiklejohn, 2002). Boelhouwers and Meiklejohn (2002) have suggested that the focus of southern African research has tended to be qualitative and lacking in scientific rigour. To this end, the current research has sought to undertake a detailed scientific approach, which will hopefully set the scene for further research undertaken on possible Quaternary deposits in southern Africa. The main concern regarding Quaternary glaciation in the Lesotho highlands has been the lack of precipitation during the LGM, given that this area falls within the summer rainfall season (Hall, 2004) and that precipitation was reduced by approximately 30% during the LGM (Partridge *et al.*, 1999; Tyson and Partridge, 2000).

This research has shown that based on temperature depressions of 6°C, as indicated by various proxy data in southern Africa, glaciers could have existed at the study sites, provided higher precipitation rates occurred than previously suggested. The lack of any proxy data indicating palaeoprecipitation in the high Drakensberg has led to using palaeotemperature data instead, as these are more widespread for southern Africa and are considered to be more robust. It is suggested that precipitation values derived for the study sites based on palaeotemperature data, would provide a better indication of palaeoprecipitation during the LGM, given that glaciers could have existed, which compare well with modern analogues. The palaeoglaciers would have relied on snowblow in order to sustain their existence and it has been suggested that there was the possibility of a greater incidence of cold fronts during the LGM, which would have increased snowfall during that time (Grab and Simpson, 2000), as discussed in Chapter 1. Contemporary snowfall preferentially accumulates at the palaeoglacier sites and remains there for longer periods than elsewhere in the region.

This has been accounted for by investigating solar radiation for a south- and north-facing slope and by modelling hillshade at the study sites.

The general consensus is that southern Africa was never glaciated during the LGM and that features ascribed to past glaciation have been misinterpreted (Preston-Whyte and Tyson, 1988; Hall, 2004). However, the linear deposits presented in this research typify attributes of glacial moraines, indicating that the study areas supported glaciers which compare well with modern analogues. The evidence does not suggest widespread glaciation in the high Drakensberg, given that evidence for the Sehonghong site indicates that this area fell within the periglacial domain, which was dominated by mass wasting processes as opposed to glacial processes. The evidence for relict LGM periglacial features also negates any proposal for widespread glaciation (e.g. Linton, 1969; Nicol, 1973; Dardis and Granger, 1986; Lewis, 1988; Lewis and Hanvey, 1988; Boelhouwers, 1991, 1994, 1998; Marker, 1992; Grab, 1997a, 2000b), given that these would not have survived had the entire area been glaciated. It is therefore proposed that conditions conducive to the formation of small niche/cirque glaciers in the eastern Drakensberg were climatologically restricted to a few sites, where microtopographic (e.g. slope gradient and aspect) and particularly microclimatic factors (e.g. late-lying snow cover, snowblow and solar radiation) played a major role.

11.4 INTERNATIONAL CONTEXT

This research indicates that during the LGM, the eastern Drakensberg was a periglacial environment, in which isolated small niche/cirque glaciers occurred. The reconstructed ELAs at the study sites between 3083 m a.s.l. and 3171 m a.s.l. (Table 11.1) also contribute to the international reconstructions of former glacier ELAs during the LGM. Similar ELAs for the African continent have been reported from Morocco (between 2500 m a.s.l. and 3600 m a.s.l.) (Messerli, 1967), and slightly higher ELAs from the eastern African mountains (e.g. 3550 m a.s.l. at Mount Kenya and between 3560 m a.s.l. to 3700 m a.s.l. at Mount Elgon) (Mahaney, 1990; Kaser and Osmaston, 2002; Osmaston, 2004). If compared to similar latitudes in the Southern Hemisphere, the reconstructed Drakensberg ELAs also fall within

Table 11.1 A Summary of similar ELAs to those observed for the Drakensberg in East Africa and the southern hemisphere.

Sites	ELAs
Drakensberg	3083 m a.s.l. to 3171 m a.s.l.
Morocco	2500 m a.s.l to 3600 m a.s.l.
East Africa	3550 m a.s.l. to 3700 m a.s.l.
Argentina	3300 m a.s.l. to 3500 m a.s.l.
Tasmania	1840 m a.s.l. to 2020 m a.s.l.

the range suggested for these latitudes. It has been suggested that in the Mendoza Andes of Argentina (35°15'S), ELAs were between 3300 m a.s.l. and 3500 m a.s.l. (Espizua, 1993), whilst in Tasmania (36°S), ELAs were between 1840 m a.s.l. and 2020 m a.s.l. during the LGM (Barrows *et al.,* 2002).

The field evidence presented in Chapters 5, 6 and 7 indicates that the Tsatsa-La-Mangaung, Sekhokong Site 1 and Leqooa Valley deposits are moraines, whilst the Sekhokong Site 2 deposit is a debris flow which was initiated on deglaciation of a small glacier. The glacier reconstructions undertaken in Chapter 10 for these sites indicate that the palaeoglaciers are comparable with modern analogues. The calibrated AMS ages for the moraines indicate that these were deposited after 13 820 yrs BP at Tsatsa-La-Mangaung, after 14 700 yrs BP at Sekhokong Site 1 and after 15 390 yrs BP at Leqooa Valley. This indicates that the moraines were deposited towards the end of the last glacial period and not during the time of maximum cooling (17 500 yrs BP) (Scott, 1982).

There appears to be controversy over whether a cold period equivalent to the Younger Dryas Stadial in north-western Europe occurred in the Southern Hemisphere (Denton *et al.,* 1999; Ivy-Ochs *et al.,* 1999; Turney *et al.,* 2003; Harrison, 2004; Kaplan *et al.,* 2004; Shulmeister *et al.,* 2005). A return to colder conditions has been suggested between 13 800 yrs BP and 11 700 yrs BP for New Zealand (Hellstrom *et al.,* 1998), whilst in Chile, it is suggested that a colder period occurred between 12 000 yrs BP and 10 000 yrs BP (Heusser *et al.,* 1999) (Table 2.10). Similarly, proxy records for southern Africa indicate a return to colder conditions between 15 000 yrs BP and 13 500 yrs BP (Holmgren *et al.,* 2003) (Table 3.2), which is also similar to the ages

obtained for the linear slope deposits. However, it has been suggested that this time period was not the coldest during the LGM (Scott, 1982; Holmgren *et al.*, 2003). The minimum ages for the deposits suggest that these were deposited between 13 820 yrs BP and 15 390 yrs BP, which agrees with the colder time period suggested by Holmgren *et al.* (2003). Although these moraines were not deposited during the coldest period during the LGM, the presence of the secondary ridges at the Tsatsa-La-Mangaung, Sekhokong Site 1 and Leqooa eastern deposits, indicate that glaciers did extend further than the position of the lateral moraines. This possibly reflects the period of maximum cooling around 17 500 yrs BP (Scott, 1982; Holmgren *et al.*, 2003), with a return to colder conditions around the time of moraine formation, when sediment supply was more readily available.

11.5 FUTURE RESEARCH IN THE DRAKENSBERG REGION

This research has identified that those areas which hosted small glaciers during the Late Pleistocene are also areas supporting contemporary late-lying snow. In order to further support this statement, areas of contemporary late-lying snow cover were studied using Google Earth (available on the internet). Unfortunately the current resolution is not satisfactory for all areas of Lesotho; however this will improve over time. Thus far, two sites have been identified as having possible linear deposits of similar morphologies to those described in this research (Figure 11.1). However, these features are relatively large and any smaller features may not have been identified using Google Earth. It is thus necessary to undertake remote sensing and ground truthing of critical sites across the entire Drakensberg range, so as to better identify other similar deposits and subsequently map these.

11.6 SUMMARY

This research has developed robust criteria permitting the identification and interpretation of deposits and landforms of glacial origin in the Lesotho Highlands. This has been achieved through a range of techniques involving morphological, sedimentological and micromorphological analysis, AMS dating and glacier reconstruction modelling. Each technique was useful in determining the process origin of the various landforms in the Drakensberg; however it is the combination of these techniques which enabled the adequate interpretation of the features investigated.

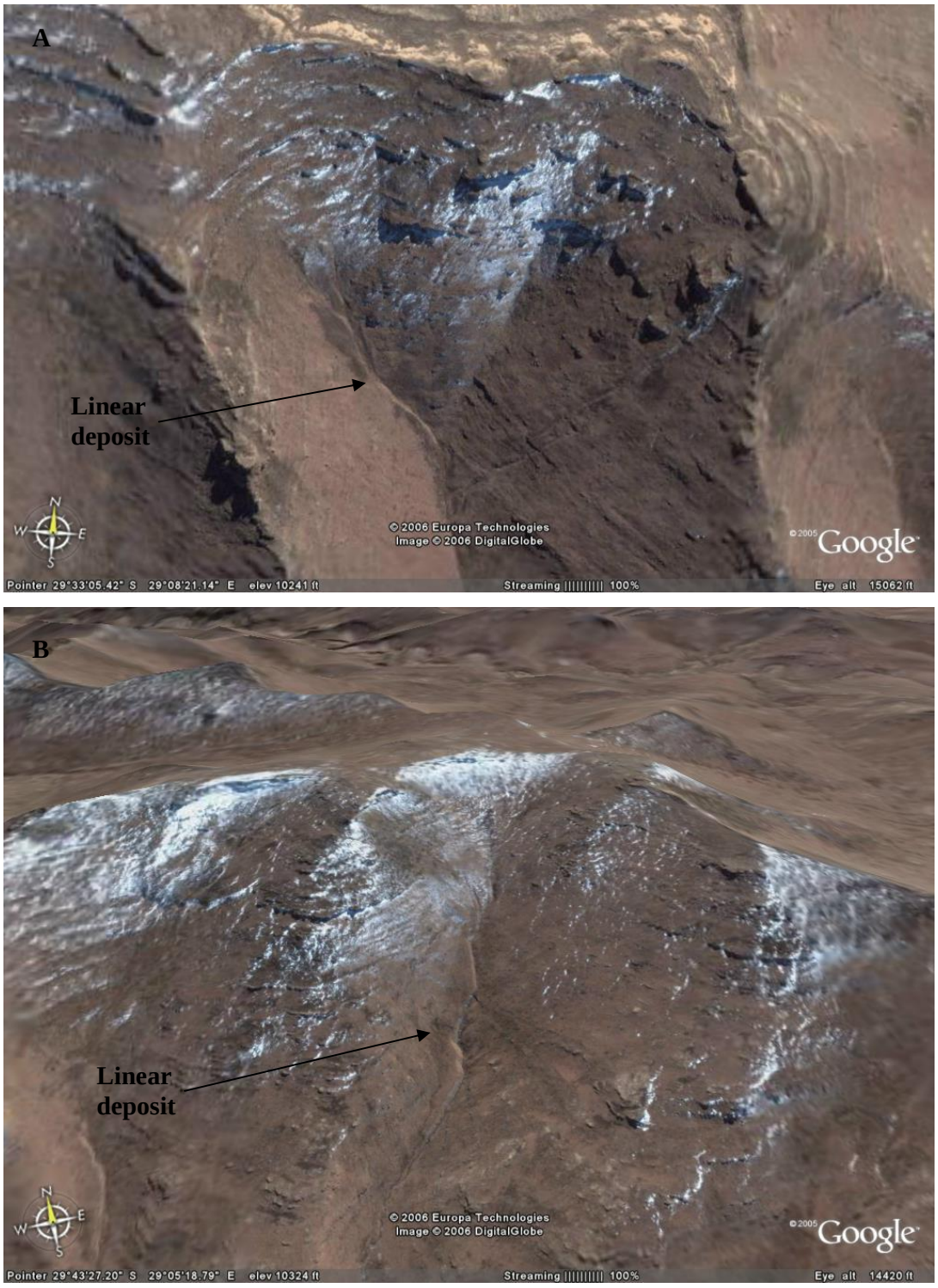


Figure 11.1 Linear deposits identified from areas of late-lying snow cover in the Drakensberg.

The use of particle shape, roundness and clast orientation alone would have been problematic given the limitations discussed in Chapter 2, however used in combination with other sedimentological properties, these proved extremely useful. A major drawback with regards to the use of micromorphological techniques is the lack of data relating to supraglacial debris. However, this research and the structures observed within the deposits will help add to the limited literature. The most useful of the techniques was the glacier reconstruction modelling, given that this could determine whether glaciers could have existed under specific climatic conditions, compared to modern analogues. Had the environmental setting not been conducive to the formation of glaciers then this would have been reflected in the thickness of glaciers and associated mass balance characteristics and a glacial origin could have been rejected.

The significance of the existence of former glaciers during the Late Quaternary with regards to palaeoclimatic reconstructions, involves new data regarding palaeoprecipitation at the study sites. The delicate relationship between temperature and precipitation in sustaining glaciers enables the determination of palaeoprecipitation at each site, based on palaeotemperature records. The fact that the study sites could have sustained glaciers indicates that precipitation values discussed in Chapter 10 were higher than at present. Despite Hall's (2004) conclusion that '*no detail can be given as to glacier distribution as, simply none exists'* (p338) for southern Africa, this research has identified the occurrence of five marginal moraines, which indicate that small glaciers did exist in the south-eastern Drakensberg. This study has therefore contributed to the Quaternary evolution of the Drakensberg landscape by identifying that although the environment was predominantly periglacial during the LGM, small glaciers were also present at certain sites.

The relative lack of glacial deposits in the high Drakensberg may be explained by the relatively short time-span of glacier occupancy in the region. Glaciers would only have developed in areas which were topographically conducive, as explained in previous sections, and would have only existed for a few thousand years. The lack of evidence for pronounced cirque glaciation is obvious for this reason, given that minor modification of the landscape by cirque glaciers occurs after approximately 70 000 yrs of occupancy, whilst sharp ridgelines and pronounced cirques become established

after approximately 200 000 yrs of occupancy (Kirkbride and Matthews, 1997). There is still much scope for future Quaternary research in the high Drakensberg, particularly to further quantify the occurrence of glaciers during the LGM and reconstruct the palaeoprecipitation at that time. The identification of similar deposits through the use of contemporary late-lying snow patches as identified in Figure 11.1 suggests that further deposits occur within this region, and which may indicate the presence of further small palaeoglaciers. It is therefore imperative that similar deposits are identified and analysed in order to further quantify the distribution of small niche/cirque glaciers during the LGM.

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