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Fine-Scale Variations of Near-Surface-Temperature Lapse Rates in the High Drakensberg Escarpment, South Africa: Environmental Implications

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Abstract

The objective of this study is to determine fine-scale spatial and temporal trends in screen temperature, lapse rates, and inversions associated with the Great African Escarpment wall. Particular attention is focused on the characteristics of near-surface-temperature lapse rates, their variability associated with particular synoptic conditions, and environmental implications such as for modeling climate projections in topographic- and climatic-specific high mountain regions. Hourly temperature data were logged on northeast- and southeast-facing slopes over 12 months in 2001, in the high Drakensberg Escarpment of southern Africa, at a range of elevations including: 2650, 2900, and 3200 m a.s.l. High spatial and temporal variations in lapse rates are recorded and illustrated. A relatively low mean annual lapse rate of $-0.42\,{^\circ}\text{C}\,100\,\text{m}^{-1}$ between 2650 and 2900 m a.s.l. doubles to $-0.84\,{^\circ}\text{C}\,100\,\text{m}^{-1}$ between 2900 and 3200 m a.s.l. The frequent westerlies and tropical temperate troughs provide for high lapse rates ($-0.91\, {^\circ}\text{C}\,100\,\text{m}^{-1}$), which are likely responsible for the relatively high annual average lapse rate of $-0.63\, {^\circ}\text{C}\,100\,\text{m}^{-1}$. Quantifying the fine-scale temperature trends has important implications for better understanding site-specific paleoenvironmental signatures and for projecting future climate scenarios and associated bio- and geosystem responses.

Introduction

Mountain climates are amongst the most complex and variable over short spatial scales. However, given the lack of climate information for many mountain environments worldwide, it is difficult to predict what impacts global and regional warming might have on local mountain climates and their environments (Pepin et al., 1999). It has been argued that air temperature is the single most important attribute of mountain climates (Barry, 1992), and previous studies have demonstrated that large variations in air temperature lapse rates add to the complexity of temperature patterns across mountainous terrain (Pepin et al., 1999; Rolland, 2003; Marshall et al., 2007; Holden et al., 2011). Many studies examining lapse rates in mountain environments are typically at a regional scale (>10 km) (e.g. Fang and Yoda, 1988; Pepin, 2001; Blandford et al., 2008), yet it is argued that finer-scale (<10 km) analyses and modeling at the local valley scale may provide for smaller temperature prediction errors (e.g. Bolstad et al., 1998; Pagés and Miró, 2010). Lapse rates in the world’s major escarpments, which represent rock walls extending several hundred meters in height and hundreds of kilometers in length, are perhaps least well understood. Yet, major escarpments such as that along eastern southern Africa have considerable influence on regional climates by contributing towards orographic uplift, by blocking circulation and the advection of coastal airflow towards the central interior, and by generating airflow (mountain winds) that influences climates in surrounding regions. This paper specifically focuses on gaining a better understanding of the fine-scale spatio-temporal characteristics of screen (i.e. near-surface) temperature, lapse rates, and inversions within a topographic cutback (steep erosional valley head) situated along the Great Drakensberg Escarpment of southern Africa, and in so doing, aims to highlight important implications for establishing site-specific thermal dynamics in alpine escarpment regions, particularly in the context of global change scenarios.

Although several studies have discussed climate attributes such as air flow (Freiman et al., 1998), precipitation (Sene et al., 1998; Nel and Sumner, 2005, 2008) and temperatures (Killick, 1978a; Grab, 1997; Nel and Sumner, 2008) for the Drakensberg escarpment, there are no published records on fine-scale temperature lapse rates and how these might vary within distinct altitudinal zones and on opposing slope aspects (northeast vs. southeast) as one approaches the escarpment summit. Equally, it would appear that little, if any, work has focused on surface temperature lapse rates for other major global escarpments such as the Serra do Mar (Brazil), Western Ghats and Deccan Traps (India), Great Escarpment of eastern Australia, and Simien Mountain escarpment (Ethiopia). The objective is thus to address such knowledge gaps by providing an example of how a major escarpment wall might spatially influence air temperature. It must be acknowledged that such fine-scale temperature trends and lapse rates will vary over small spatial scales, and thus the findings cannot necessarily be extrapolated for all escarpments, but rather be used as an example to highlight the complexity of temperature and lapse rate variability in such topographic settings. In addition, the amplitude of climate change at lower altitudes may not represent those at higher altitudes, as surface temperature lapse rates (and indeed other climate parameters) are likely to vary in accordance with synoptic weather events, each of which is associated with particular moisture, airflow, radiation, and temperature patterns (cf. Pepin et al., 1999; Rolland, 2003; Hope, 2006; Blandford et al., 2008). Thus, should...
climate models project a higher future frequency for particular synoptic events, it may be possible, on such a basis, to model likely temperature changes at higher altitudes where mean contemporary temperature lapse rates are known for such events. This may help improve site-specific projections of how regolith, biota, and water resources may respond to current and future climate change in the high escarpment zone (cf. Holden et al., 2011).

**Methods**

**STUDY AREA AND PREVIOUS WORK**

The Great Escarpment of southern Africa attains its maximum altitude for about 280 km along the KwaZulu-Natal/LeSotho border where it averages ca. 2900–3000 m a.s.l. (Fig. 1). The high escarpment zone is an important “water tower” in southern Africa, where annual precipitation (estimated between 800 and 1500 mm yr⁻¹; Sene et al., 1998; Nel and Sumner, 2008) exceeds evaporation (Zuncel, 2003). Floristically, the high Drakensberg escarpment belongs to the Drakensberg Alpine Centre (DAC), a composite of high-altitude (>1700 m a.s.l.) enclaves, which hosts rare and endemic flora with high biodiversity (Carbutt and Edwards, 2004, 2006). Although the DAC is a grassland biome, it has previously been classified into three altitudinal belts according to climax communities: *Podocarpus latifolius* forest (montane belt) between 1280 and 1830 m a.s.l., *Passerina-Philippia-Widringtonia* fynbos (subalpine belt) between 1830 and 2865 m a.s.l., and *Erica-Helichrysum* heath (alpine belt) above 2865 m a.s.l. (Killick, 1963). However, the lower limit of the alpine belt remains poorly defined, but is suggested to be at ca. 2740 m (Killick, 1978b). Given the natural and cultural heritage of the Drakensberg escarpment and adjoining foothills, the region was declared a World Heritage Site in 2000 and also falls within the Maloti-Drakensberg Transfrontier Project (MDTP) (Fig. 1).

Killick (1978b) provided the first detailed air temperature assessment for the high Lesotho plateau (2400–3050 m a.s.l.), based on data obtained for the period 1966–1976. These data were subsequently used in an expanded study that compared altitudinal trends in temperature, from Butha-Buthe in the montane belt (1768 m a.s.l.) to Letseng-la-Draai in the alpine belt (3050 m a.s.l.) (Grab, 1997). The study examined lapse rates at a regional scale and calculated exceptionally steep lapse rates in the mid- to upper-subalpine belt, averaging −1.9 °C 100 m⁻¹, and then a substantially reduced lapse rate of −0.3 °C 100 m⁻¹ from the upper subalpine belt to the alpine belt at 3050 m a.s.l. (Grab, 1997). Given that this previous study examined lapse rates over a region of ca. 4200 km², such documented lapse rates highlight the likely influence of topographic position, in addition to altitude, and thus provide limited indications on lapse rates at local spatial scales (i.e. within an individual valley or along mountain slopes). More recently, Nel and Sumner (2008) described spatial attributes of rainfall and temperature, yet only used two weather stations set several kilometers apart. Thus, the current study is the most detailed fine-scale spatial and temporal analysis of near-surface temperature along the Great Escarpment of southern Africa.

**DATA, METHODS, AND CHALLENGES**

Six screen temperature stations were installed in the Nhlangeni cutback (29°29’S; 29°18’E, Figs. 1 and 2), with the aim to examine fine-scale near-surface-temperature lapse rates (+1.5 m above the ground surface) from the upper subalpine belt (2650 m a.s.l.) to the alpine belt (3200 m a.s.l.). Monitoring commenced from September 1999 until October 2003, after which the equipment was stolen. Stations were positioned on northeast- (NE) and southeast- (SE) facing cutback slopes along the main escarpment at 2650, 2900, and 3200 m a.s.l. (Figs. 1 and 2). The altitudinal zone between 2650 and 2900 m a.s.l. represents the lower portion of the escarpment wall, whilst the zone between 2900 and 3200 m a.s.l. represents the upper portion of the escarpment, and environmentally constitutes alpine and periglacial attributes. Screen temperatures were also recorded on the escarpment summit plateau at Sani Top (2874 m a.s.l.), where instrumentation is safe from vandalism or theft (Fig. 1, part d). The site is ca. 200 m west of the escarpment edge and 10 km south of the Nhlangeni cutback. The summit plateau temperatures immediately to the west of the escarpment may thus be compared to those at a similar altitude along the east-facing escarpment zone. Screen temperatures were logged at 1 h intervals using calibrated Tinytag temperature loggers with a resolution of ±0.2 °C. The pencil probe thermistors were housed in radiation shields set 1.5 m above ground. Wind data were provided by the Risø National Laboratory, Denmark, which had set up an experimental wind monitoring program at Sani Top in 2001, at 2870 m a.s.l., ca. 800 m west of the escarpment edge (Fig. 1, part d). Mean wind speed and direction were logged at 10 min intervals, 9 m above ground.

Data screening revealed considerable data errors due to logger failures. Given the remoteness of the study site, it was only feasible to download data every 6 to 8 months. Nevertheless, all temperature stations recorded 100% data retrieval during 2001, which also coincided with the availability of wind data from the Sani plateau. This study hence uses the 2001 data to scrutinize fine-scale trends in temperature and lapse rates. According to Grab (2004), the year 2001 had a mean annual air temperature of 0.3 °C below the 20-yr mean, and winter temperatures (May–August) coincided exactly with the 20-yr mean. However, the year had an exceptionally cold January (ca. 1.2 °C below normal), thus apart from this month, the year of analysis was climatically “typical” in the context of other years. Although 12 months’ data are temporally limiting, these should be sufficient to gain new insight on temperature variability and lapse rates near the upper escarpment edge. It is also acknowledged that wind characteristics recorded at the Sani Top plateau, 10 km from the Nhlangeni cutback, do not necessarily correspond with local airflow exposed along the east-facing escarpment wall. Nevertheless, the wind data are likely to provide a valuable indication of the relative temporal patterns of wind direction and strength in the region. Unfortunately the timing, depth, and distribution of snow cover could not be established; however, it is acknowledged that snow cover would have had an influence on the results during the short periods of time when it did occur.

In this paper, near-surface lapse rates are expressed as °C 100 m⁻¹ rather than °C km⁻¹, given the fine-scale approach, with only a 550 m altitudinal range being investigated.

**Results**

**TEMPERATURE TRENDS AND FROST DAYS IN 2001**

The mean temperature for Sani Top (2874 m a.s.l.) to the east of the escarpment (Fig. 1) was 6.1 °C, thus considerably lower than those at a comparable altitude (2900 m a.s.l.) in the Nhlangeni...
FIGURE 1. (A) The study area in southern Africa, (B) site localities in eastern Lesotho, (C) Nhlangeni cutback, and (D) Sani Top plateau.
The Nhlangeni cutback (av: 7.5 °C for SE-facing aspect; 7.9 °C for NE-facing aspect) (Table 1). Whilst mean annual maximum temperatures at Sani Top are within 0.6 °C of those measured in the Nhlangeni cutback at 2900 m a.s.l., mean minimum temperatures on the plateau are substantially lower (by 3.4–3.6 °C) than those measured at a comparable altitude in the Nhlangeni cutback. Such pronounced temperature differences between the plateau and escarpment cutback demonstrate, in particular, the strong topo-climatic influence on nocturnal radiative cooling and cold air drainage on the plateau. In contrast, cold air drainage or pooling is negligible in the Nhlangeni cutback; thus, the measured lapse rates are broadly representative of large-scale free air effects along the escarpment.

The 2001 temperature data for the Nhlangeni cutback indicate mean annual temperatures that vary from 9.0 °C at 2650 m a.s.l. (NE-facing) to 4.9 °C at 3200 m a.s.l. (SE-facing) (Table 2). Mean summer temperatures vary from 11.7 °C at 2650 m a.s.l. (NE-facing) to 8.6 °C at 3200 m a.s.l. (SE-facing), whilst those in winter range from 5.2 °C at 2650 m a.s.l. (NE-facing) to 0.5 °C at 3200 m a.s.l. (SE-facing). Thus, temperature contrasts between the lower and upper parts of the escarpment cutback are considerably higher during winter (4.7 °C) than summer (3.1 °C). Extreme temperatures during 2001 varied from 23.7 °C at 2650 m a.s.l. (SE-facing) to –11.7 °C at 3200 m a.s.l. (NE-facing).

Typically, SE-facing slopes are on average 0.4–0.5 °C cooler...
Table 1:
Comparison of mean monthly temperature variables during 2001 for Sani Top (2874 m a.s.l.) and the Nhlangeni (Nhl) cutback (2900 m a.s.l.).

<table>
<thead>
<tr>
<th></th>
<th>Sani</th>
<th>Nhl SE</th>
<th>Nhl NE</th>
</tr>
</thead>
<tbody>
<tr>
<td>Jan</td>
<td>10</td>
<td>11.2</td>
<td>11.1</td>
</tr>
<tr>
<td>Feb</td>
<td>9</td>
<td>10.2</td>
<td>10.3</td>
</tr>
<tr>
<td>Mar</td>
<td>9.2</td>
<td>10.6</td>
<td>10.8</td>
</tr>
<tr>
<td>Apr</td>
<td>6.3</td>
<td>6.8</td>
<td>7.1</td>
</tr>
<tr>
<td>May</td>
<td>4.2</td>
<td>5.8</td>
<td>6.7</td>
</tr>
<tr>
<td>Jun</td>
<td>1.1</td>
<td>3.2</td>
<td>4.2</td>
</tr>
<tr>
<td>Jul</td>
<td>0.4</td>
<td>1.7</td>
<td>2.7</td>
</tr>
<tr>
<td>Aug</td>
<td>3.5</td>
<td>5.1</td>
<td>5.9</td>
</tr>
<tr>
<td>Sep</td>
<td>4.6</td>
<td>6.1</td>
<td>6.4</td>
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<tr>
<td>Oct</td>
<td>8.1</td>
<td>9.3</td>
<td>9.4</td>
</tr>
<tr>
<td>Nov</td>
<td>8.5</td>
<td>9.6</td>
<td>9.5</td>
</tr>
<tr>
<td>Dec</td>
<td>9.3</td>
<td>10.5</td>
<td>10.3</td>
</tr>
<tr>
<td>Year</td>
<td>6.1</td>
<td>7.5</td>
<td>7.9</td>
</tr>
</tbody>
</table>

Table 2:
Statistical summary of results based on temperature monitoring for various aspects and altitudes in the Nhlangeni cutback (Jan–Dec 2001).

<table>
<thead>
<tr>
<th>Temperature</th>
<th>2650SE</th>
<th>2650NE</th>
<th>2900SE</th>
<th>2900NE</th>
<th>3200SE</th>
<th>3200NE</th>
<th>Sani plateau 2874</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mean</td>
<td>8.5</td>
<td>9</td>
<td>7.5</td>
<td>7.9</td>
<td>4.9</td>
<td>5.4</td>
<td>6.1</td>
</tr>
<tr>
<td>Abs. max</td>
<td>23.7</td>
<td>23</td>
<td>21.6</td>
<td>21.3</td>
<td>19.9</td>
<td>21.7</td>
<td>21.1</td>
</tr>
<tr>
<td>Abs. min</td>
<td>−7.4</td>
<td>−7.6</td>
<td>−8.5</td>
<td>−8.1</td>
<td>−11.7</td>
<td>−12</td>
<td>−12.8</td>
</tr>
<tr>
<td>Frost days</td>
<td>38</td>
<td>36</td>
<td>74</td>
<td>41</td>
<td>160</td>
<td>120</td>
<td>119</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Lapse rates (°C/100 m)</th>
<th>2650–2900SE</th>
<th>2650–2900NE</th>
<th>2900–3200SE</th>
<th>2900–3200NE</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mean</td>
<td>0.39</td>
<td>0.44</td>
<td>0.86</td>
<td>0.81</td>
</tr>
<tr>
<td>Abs. max</td>
<td>2.88</td>
<td>2.72</td>
<td>3.63</td>
<td>3.16</td>
</tr>
<tr>
<td>Inversions (°C/100 m)</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Mean</td>
<td>−0.46</td>
<td>−0.51</td>
<td>−0.39</td>
<td>−0.45</td>
</tr>
<tr>
<td>Abs. max</td>
<td>−3.16</td>
<td>−3.4</td>
<td>−2.68</td>
<td>−2.6</td>
</tr>
<tr>
<td>Number (hrs)</td>
<td>663</td>
<td>806</td>
<td>500</td>
<td>954</td>
</tr>
</tbody>
</table>
14 frost days/100 m (lower cutback) to 28 frost days/100 m (upper cutback). The trend is much less pronounced on NE-facing slopes where there is an advance of only 2 frost days/100 m in the lower cutback, but increases to 26 frost days/100 m in the upper cutback.

**LAPSE RATES IN 2001**

Although previous international work has demonstrated considerable variation of environmental lapse rates in mountain environments, and high coefficients of variation between sites within given mountain regions (De Scally, 1997), mean values of between $-0.5$ and $-0.65 \, ^\circ C \, 100 \, m^{-1}$ are regularly quoted (Burk and Stuiver, 1981; Barry and Chorley, 1987; Barry, 1992; Meyer, 1992; Rolland, 2003). The Nhalenji cutback results indicate considerable spatial and temporal variations in lapse rates (Table 2: Figs. 3 and 4). Despite a relatively low mean annual lapse rate ($-0.42 \, ^\circ C \, 100 \, m^{-1}$) between 2650 and 2900 m a.s.l., this doubles to $-0.84 \, ^\circ C \, 100 \, m^{-1}$ in the upper escarpment zone between 2900 and 3200 m a.s.l., and averages $-0.63 \, ^\circ C \, 100 \, m^{-1}$ between 2650 and 3200 m a.s.l. Maximum lapse rates vary from $-2.88 \, ^\circ C \, 100 \, m^{-1}$ between 2650 and 2900 m a.s.l. to $-3.63 \, ^\circ C \, 100 \, m^{-1}$ between 2900 and 3200 m a.s.l.

Data show a pronounced seasonal trend in mean monthly variations of lapse rates between 2650 and 3200 m a.s.l., being lowest during winter (Nov–Feb) ($-0.54 \, ^\circ C \, 100 \, m^{-1}$) and highest during spring (Aug–Oct) ($-0.76 \, ^\circ C \, 100 \, m^{-1}$) (Figs. 3 and 4). Lapse rates during autumn and winter (March–July) average $-0.62 \, ^\circ C \, 100 \, m^{-1}$. The results are in strong contrast to several other studies that have reported higher summer than winter lapse rates in other mountain regions (e.g. Ozenda, 1985; Hutchinson, 1991; Rolland, 2003). The annual maximum monthly variance of lapse rates on NE- and SE-facing slopes between 2650 and 2900 m a.s.l. (50% and 52%, respectively) is similar to that between 2900 and 3200 m a.s.l. (52% and 47%, respectively).

Mean $T_{\text{min}}$ lapse rates are considerably higher ($-0.66 \, ^\circ C \, 100 \, m^{-1}$) than those for mean $T_{\text{max}}$ ($-0.53 \, ^\circ C \, 100 \, m^{-1}$) between 2650 and 3200 m a.s.l. (Fig. 4). Mean annual $T_{\text{max}}$ lapse rates are higher between 2650 and 2900 m a.s.l. ($-0.60 \, ^\circ C \, 100 \, m^{-1}$) than between 2900 and 3200 m a.s.l. ($-0.47 \, ^\circ C \, 100 \, m^{-1}$); however, this trend is reversed for $T_{\text{min}}$, for which mean annual lapse rates increase from $-0.32 \, ^\circ C \, 100 \, m^{-1}$ between 2650 and 2900 m a.s.l. to $-1.00 \, ^\circ C \, 100 \, m^{-1}$ between 2900 and 3200 m a.s.l. Further to this, whilst mean annual $T_{\text{max}}$ lapse rates between 2650 and 2900 m a.s.l. on NE- and SE-facing slopes only differ fractionally ($-0.60$ and $-0.59 \, ^\circ C \, 100 \, m^{-1}$), those between 2900 and 3200 m a.s.l. differ substantially between NE- and SE-facing slopes ($-0.14$ and $-0.80 \, ^\circ C \, 100 \, m^{-1}$), respectively. The most pronounced spatio-temporal contrast in $T_{\text{max}}$ lapse rates in the upper escarpment zone (2900–3200 m a.s.l.) occurs between NE-facing slopes during the warm period (Dec–March) when mean conditions reflect a slight temperature inversion ($+0.04 \, ^\circ C \, 100 \, m^{-1}$) and SE-facing slopes during August/September when lapse rates average $-1.29 \, ^\circ C \, 100 \, m^{-1}$. Mean $T_{\text{min}}$ lapse rates are considerably suppressed at lower altitudes (2650–2900 m), but somewhat higher during the cold season (May–Aug) ($-0.38 \, ^\circ C \, 100 \, m^{-1}$) than during the warm season (Dec–March) ($-0.27 \, ^\circ C \, 100 \, m^{-1}$). In contrast, $T_{\text{min}}$ lapse rates in the upper escarpment zone remain relatively high throughout the year ($\geq -0.6 \, ^\circ C \, 100 \, m^{-1}$) but peak during the cold season when they average $-1.21 \, ^\circ C \, 100 \, m^{-1}$.

Mean nocturnal (18:00–06:00) lapse rates are considerably lower ($-0.33 \, ^\circ C \, 100 \, m^{-1}$) than those during the day (07:00–17:00) ($-0.5 \, ^\circ C \, 100 \, m^{-1}$) in the lower escarpment zone; this trend is reversed in the upper escarpment zone, which reflects stronger nocturnal ($-1.04 \, ^\circ C \, 100 \, m^{-1}$) than day-time ($-0.63 \, ^\circ C \, 100 \, m^{-1}$) lapse rates (Fig. 5). In other upland valleys such as in the English Pennines, lapse rates are reportedly lowest during midday for both upper and lower slopes (Pepin et al., 1999), and thus the current findings are likely to be influenced by the macro-topography of the escarpment wall and associated insolation received. Lapse rates on both NE- and SE-facing slopes between 2650 and 3200 m a.s.l. are steepest between 04:00 and 07:00. Lapse rates drop rapidly between 06:00 and 09:00 on NE-facing slopes; those on SE-facing slopes drop much more gradually until mid-afternoon (15:00) when they reach their lowest values ($-0.6 \, ^\circ C \, 100 \, m^{-1}$).

**TEMPERATURE INVERSIONS IN 2001**

Temperature inversions are somewhat stronger and more frequent on NE-facing slopes than SE-facing slopes (Fig. 6). On NE-facing slopes they annually occur from 9.2% (806 hrs) to 10.9% (954 hrs) of time for the lower and upper escarpment zones, respectively, and average from $+0.51$ to $+0.45 \, ^\circ C \, 100 \, m^{-1}$, respectively. In contrast, on SE-facing slopes inversions occur from 7.6% (663 hrs) to 5.7% (500 hrs) of time for the lower and upper escarpment zones, respectively, and average from $+0.46$ to $+0.39 \, ^\circ C \, 100 \, m^{-1}$, respectively. The most notable characteristic is that the frequency of inversions in the upper escarpment NE-facing slope is almost double that for the opposing SE-facing slope. On average, inversions strengthen from the upper towards the lower escarpment zones, as is also the case for absolute values which are $+2.7$ and $+3.4 \, ^\circ C \, 100 \, m^{-1}$, respectively. Inversions are most frequent during mid-summer (Jan/Feb) and mid-winter (June/July) when they occur on average from 12.6% to 11.2% of time, respectively, and are lowest during early spring (Aug) and mid-autumn (April) when they occur on average from 4.9% to 5.7% of time, respectively. In the lower escarpment zone, inversions are slightly more frequent during nocturnal hours (19:00–06:00) (9.1% of days) than during diurnal hours (07:00–18:00) (8.0% of days). However, in the upper escarpment zone, nocturnal inversions are infrequent (1.8% of days) whilst diurnal inversions are most common (14.9% of days). The frequent daytime summer inversions are, amongst other possible reasons, likely due to low level clouds at lower elevations along the Great Escarpment.

The hourly frequency distribution of temperature inversions varies considerably with seasonality for both the lower and upper escarpment zones (Fig. 7). In the lower escarpment zone, the hourly frequency of mid-summer (January) inversions is relatively similar both in direction and deviation between SE-facing (s.d. = 2.7) and NE-facing (s.d. = 2.5) exposures, with a pronounced absence between 06:00 and 07:00. The greatest daily contrast in the hourly inversion frequencies between SE- and NE-facing exposures is during winter (June) when the standard deviation on SE-facing exposures is 1.8 whilst that on NE-facing exposures is 7.3. On SE-facing exposures the frequency of inversions during this time of year is highest during the early morning (03:00–04:00; 5% of days).
FIGURE 3. Mean daily lapse rates for various aspects and altitudes (Jan–Dec 2001).
FIGURE 4. Mean monthly $T_{\text{max}}$, $T_{\text{min}}$, $T_{\text{mean}}$ lapse rates for various aspects and altitudes (Jan–Dec 2001).
and at 19:00 (6% of days), whilst on NE-facing exposures it is highest between 10:00 and 13:00 (mean frequency = 19.8% of days). The frequency and hourly deviation of temperature inversions in the upper escarpment zone are inter-seasonally more distinct and consistent than in the lower escarpment zone. Temperature inversions are negligible during nocturnal hours throughout the year. A relatively high frequency of diurnal temperature inversions are measured during all seasons for upper escarpment NE-facing exposures. The highest frequencies are during mid-mornings in mid-summer (19% of days at 10:00) and in winter (21% of days at 09:00), whilst during the intermediate seasons they peak slightly later at between 10:00 and 12:00, and occur on up to 12% of days in autumn (at 12:00) and 10% of days in early spring (at 10:00).

SYNOPTIC CONDITIONS AND TEMPERATURE LAPSE RATES IN 2001

Although relatively few studies have investigated the influence of synoptic variability on lapse rates, it is an important factor to consider given its control on radiative exchange and mixing (Pepin and Kidd, 2006). Typical synoptic patterns that impact the annual climate of the Drakensberg escarpment, as classified by Tyson and Preston-Whyte (2000), include tropical temperature troughs, east coast low pressure systems, and southeast coast ridging high pressure systems that are most dominant during warmer seasons, whilst westerlies, high pressure systems over the interior, cold fronts, and ridging high pressure systems behind cold fronts are most dominant during the cooler seasons. With the aid of daily synoptic charts from 2001, provided by the South African Weather Service, it is possible to determine the temporal occurrence of such synoptic events and ascertain their influence on temperature lapse rates along the Great Escarpment. The charts indicate conditions at 14:00 each day, providing a description of the synoptic patterns together with prevailing surface pressure, wind direction/speed, and the location of cold/warm fronts, coastal low pressure systems, etc.

Synoptic conditions that produce mean temperature lapse rates between 2650 and 3200 m a.s.l., which most closely reflect the annual mean lapse rate of \(-0.63\, ^\circ C/100\, m\), are associated with cold fronts (mean = \(-0.59\, ^\circ C/100\, m\)) (Table 3). Mean lapse rates connected with southeast coast ridging high pressure systems and ridging high pressure systems behind cold fronts result in particularly weak lapse rates (\(-0.08\) and \(-0.2\, ^\circ C/100\, m\), respectively), whilst those associated with tropical temperature troughs, east coast lows, and westerlies provide for the strongest lapse rates (\(-0.71, -0.83,\) and \(-0.91\, ^\circ C/100\, m\), respectively). Airflow connected with the southeast coast ridging high and ridging high behind cold fronts is humid and southeasterly (mean wind vector: \(142^\circ -145^\circ\); Fig. 8), with consequent cloud development along the Great Escarpment. However, such cloud typically thins towards the upper escarpment zone, where insolation levels increase, thus accounting for weak lapse rates. This inference is supported when examining changes in lapse rates from the lower to upper escarpment zones. On SE-facing exposures, mean lapse rates decrease from \(-0.17\) to \(-0.06\, ^\circ C/100\, m\), whilst those on NE-facing exposures decrease from \(-0.24\) to \(+0.12\, ^\circ C/100\, m\) (i.e. mean conditions reflecting a temperature inversion). The strongest lapse rates are associated with westerly to northwesterly airflow (mean wind vectors ranging from \(262^\circ\) to \(292^\circ\); Fig. 8), and particularly in the case of westerlies, are atypical to the European Föhn conditions with pronounced descending gradient winds which mask out anabatic and catabatic valley winds. Notably, \(-61\%\) of airflow during 2001 was from a westerly to northwesterly direction (Fig. 8).
FIGURE 6. Number of hours with temperature inversions per month (Jan-Dec 2001) and percentage of days experiencing temperature inversions during particular hours of the day.
Lapse rates are most consistent between the lower (−0.21 °C 100 m⁻¹) and upper (−0.19 °C 100 m⁻¹) escarpment zones during the phase of ridging high pressure systems behind cold fronts. However, the westerly airflow connected with the overpass of cold fronts on their own account for higher mean lapse rates, which in particular are strengthened on SE-facing exposures from the lower (−0.48 °C 100 m⁻¹) to upper (−0.82 °C 100 m⁻¹) escarpment zones. The greatest strengthening of lapse rates (by 51%) from the lower to upper escarpment zones are associated with tropical temperature troughs and high pressure systems over the interior, which coincidently are synoptic events with the lowest mean wind speeds (3.7 and 3.6 m s⁻¹, respectively) (Table 3). Strong
TABLE 3
Mean wind direction, wind speed, and lapse rates for various aspects and altitudes during particular synoptic conditions (Jan–Dec 2001).

<table>
<thead>
<tr>
<th>Wind dir. (m s(^{-1}))</th>
<th>Wind speed (m s(^{-1}))</th>
<th>2650–2900 (SE-facing)</th>
<th>2650–2900 (NE-facing)</th>
<th>2900–3200 (SE-facing)</th>
<th>2900–3200 (NE-facing)</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Cold fronts and ridging high (n = 11)</strong></td>
<td></td>
<td></td>
<td></td>
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</tr>
<tr>
<td>Av.</td>
<td>142°</td>
<td>3.9</td>
<td>0.24</td>
<td>0.17</td>
<td>0.19</td>
</tr>
<tr>
<td>s.d.</td>
<td>27°</td>
<td>1.7</td>
<td>0.37</td>
<td>0.33</td>
<td>0.46</td>
</tr>
<tr>
<td><strong>Cold fronts (n = 9)</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Av.</td>
<td>278°</td>
<td>7.9</td>
<td>0.48</td>
<td>0.57</td>
<td>0.82</td>
</tr>
<tr>
<td>s.d.</td>
<td>21°</td>
<td>3</td>
<td>0.29</td>
<td>0.39</td>
<td>0.54</td>
</tr>
<tr>
<td><strong>Tropical Temp trough (n = 14)</strong></td>
<td></td>
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<tr>
<td>Av.</td>
<td>262°</td>
<td>3.7</td>
<td>0.43</td>
<td>0.5</td>
<td>0.95</td>
</tr>
<tr>
<td>s.d.</td>
<td>40°</td>
<td>2.1</td>
<td>0.4</td>
<td>0.25</td>
<td>0.22</td>
</tr>
<tr>
<td><strong>East coast low (n = 9)</strong></td>
<td></td>
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<tr>
<td>Av.</td>
<td>288°</td>
<td>7.5</td>
<td>0.56</td>
<td>0.66</td>
<td>1.15</td>
</tr>
<tr>
<td>s.d.</td>
<td>9°</td>
<td>3.1</td>
<td>0.31</td>
<td>0.25</td>
<td>0.2</td>
</tr>
<tr>
<td><strong>SE coast ridging High (n = 6)</strong></td>
<td></td>
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<tr>
<td>Av.</td>
<td>145°</td>
<td>6</td>
<td>0.17</td>
<td>0.24</td>
<td>0.06</td>
</tr>
<tr>
<td>s.d.</td>
<td>35°</td>
<td>1.8</td>
<td>0.41</td>
<td>0.2</td>
<td>0.26</td>
</tr>
<tr>
<td><strong>Westerlies: pre–cold front (n = 12)</strong></td>
<td></td>
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</tr>
<tr>
<td>Av.</td>
<td>292°</td>
<td>11.4</td>
<td>0.68</td>
<td>0.68</td>
<td>1.25</td>
</tr>
<tr>
<td>s.d.</td>
<td>9°</td>
<td>4.8</td>
<td>0.37</td>
<td>0.33</td>
<td>0.16</td>
</tr>
<tr>
<td><strong>High pressure over interior (n = 8)</strong></td>
<td></td>
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<tr>
<td>Av.</td>
<td>163°</td>
<td>3.6</td>
<td>0.42</td>
<td>0.26</td>
<td>0.72</td>
</tr>
<tr>
<td>s.d.</td>
<td>72°</td>
<td>1.2</td>
<td>0.2</td>
<td>0.36</td>
<td>0.53</td>
</tr>
</tbody>
</table>

northwesterly airflow may be associated with east coast lows (mean wind speed = 7.5 m s\(^{-1}\)) or approaching low pressure systems (cold fronts) to the south (mean wind speed = 11.4 m s\(^{-1}\)); such situations account for the second most strengthening of lapse rates from the lower to upper escarpment zones (by 42% and 40%, respectively).

Lapse rates double from the lower to upper escarpment zones during the presence of tropical temperature troughs for both SE-facing (by 56%) and NE-facing (by 48%) exposures, from an average of −0.46 °C 100 m \(^{-1}\) in the lower escarpment zone to −0.96 °C 100 m \(^{-1}\) in the upper escarpment zone.

Discussion and Conclusions

This paper provides some detailed fine-scale spatio-temporal trends in temperatures, lapse rates, and inversions along one of the world’s major escarpments, namely the Drakensberg. The tremendous variability and complexity in both spatial and temporal temperature trends demonstrated in this paper are a likely function of several primary factors, including site-specific topographic controls (i.e. reception of solar radiation, shading, control on air flow), albedo (e.g. percentage vegetation, snow cover), and primary synoptic conditions (see also Pepin and Kidd, 2006), which in many instances interact or dominate over the site-specific factors. The locality of a major escarpment wall (several hundred kilometers in length and ca. 800–1200 m in vertical extent) in a subtropical region, separating maritime moist air to the east (Indian Ocean) from dry air to the west, is likely to have a major influence on the observed temperature and lapse rate trends. The typically reported nocturnal cold air drainage reducing lapse rates at night, along with steeper lapse rates during the day, in summer, or associated with enhanced cyclonic activity (low pressure and high relative vorticity) (e.g. Pepin, 2001; Blandford et al., 2008; Gillies et al., 2010; Bianco et al., 2011), are all to a considerable extent masked along the Drakensberg escarpment. This is owing to regional-scale westerly winds (particularly during the austral winter and spring) bringing dry air over and down the Great Escarpment, eradicating clouds and thus steepening lapse rates towards the upper escarpment zone. In addition, easterly airflow bringing humid air typically produces cloud cover at lower elevations along the great escarpment when higher elevations remain cloud free, thus contributing to temperature inversions during the day and steep nocturnal lapse rates from the mid- to upper escarpment zones. These findings concur with those from more tropical mountains such as Kilimanjaro (cf. Duane et al., 2008).

Findings from the Nhlangeni cutback along the Drakensberg escarpment in 2001 show a substantial steepening (doubling) of mean lapse rates with altitude, from −0.42 °C 100 m \(^{-1}\) in the lower escarpment zone (2650–2900 m a.s.l.) to −0.84 °C 100 m \(^{-1}\) in the upper escarpment zone (2900–3200 m a.s.l.). Given that the year 2001 was a relatively typical climatic year and that lapse rates associated with particular synoptic conditions are unlikely to differ across multiple years, the trends observed should be representative for at least recent times. Pepin and Losleben (2002) have similarly reported a steepening of lapse rates with increasing altitude for the Colorado Rocky Mountains in the U.S.A. It is postulated that an important factor determining the annual mean lapse rates are the frequency and duration of particular synoptic events. Westerlies and tropical temperate troughs were particularly frequent during 2001 (Table 2) and accounted for mean lapse rates of −0.91 and −0.71 °C 100 m \(^{-1}\), respectively, and are likely responsible for the relatively high annual average lapse rate. Any inter-annual vari-
FIGURE 8. Mean wind directions (top rose diagram) and the relationship between mean temperature lapse rate (°C 100 m⁻¹) and wind direction (bottom rose diagram) for the period Jan–Dec 2001.
ability is thus likely controlled by the strength and frequency of particular synoptic events. This has important implications for climate projections in mountain environments such as the Drakensberg. In most instances where mountain climate data are sparse or absent, climate projections for higher altitude environments are assumed on the basis of projections calculated from General Circulation Models (GCMs) and/or are modeled from instrumental data obtained from lower altitudes; in both cases this can lead to poor representation of the situation on the ground (see Pepin and Losleben, 2002). Consequently, utmost caution and an understanding of, for instance, how the strength and frequency of particular synoptic events might change under climate change scenarios are required. Depending on changes in projected future dominant synoptic patterns, which may either strengthen or weaken mean lapse rates, higher mountain regions may experience faster or slower warming trends than those of surrounding lower regions. In addition, the global locality (i.e. latitude, continentality, etc.) and topographic structure (morphology, primary alignment, and orientation) of mountain environments are evidently important controls that need to be considered when discussing disparate trends in mountain climates at a global scale. These factors operate synergistically to impact surface temperature lapse rates and may account for conflicting trends of climate change between areas of higher elevation or between lowland areas and adjacent high mountain regions (cf. Pepin and Lundquist, 2008; Rangwala and Miller, 2012). Whilst warming trends have intensified with an increase in altitude over Tibet and the Himalaya between 1955 and 1996 (e.g. Shrestha et al., 1999; Liu and Chen, 2000), those at higher elevations in the Colorado Rockies have experienced a cooling trend between 1952 and 1998, contrary to the situation at lower altitudes (Pepin and Losleben, 2002).

The large contrasts of temperature and lapse rate trends over small altitudinal scales (few hundred meters), such as for instance mean lapse rates varying from −0.68 to −1.14 °C 100 m−1 from the lower (2650–2900 m) to upper (2900–3200 m) escarpment zone during westerly air flow, have important implications for understanding contemporary fine-scale spatial trends in faunal and floral species distributions/successions/extinctions/invasions, etc., as well as geomorphic, pedogenic, and hydrological processes (e.g. rock weathering and associated slope stability; soil chemistry; carbon cycle, water supply, and hazards associated with snow accumulation/melt, etc.). Bioclimatic models used in complex mountainous terrain are sometimes considered too course to accommodate projected geographic shifts in species, as in most instances they are unable to account for topographically induced trends in temperature, and thus a call has been made for employing statistical downscaling techniques to better understand local patterns of climate and biosystem changes (Nogués-Bravo et al., 2007; Holden et al., 2011). Further, given that several of the world’s major escarpments consist of basalt, which weather faster than other silicates, it has been argued that they significantly influence the atmospheric CO2 budget over long temporal scales (Das et al., 2005). Surface (and associated rock) temperature trends along such escarpments are fundamental to controlling rates and intensity of weathering (Grab, 2007), and thus any future projections of weathering rates that influence CO2 budgets and rates of landscape evolution need to consider the fine-scale climate trends.

Mountains are said to represent “unique” areas for detecting climatic change and for assessing climate-related impacts (Beniston, 2003), and there is wide consensus that alpine environments are particularly sensitive to global warming given their high biodiversity, refugia for endemic biota, and representation as a geographic end-point (see also Beniston, 2006). Thus, an improved understanding of the fine-scale bio-/geoclimatic systems, such as considering local lapse rates in complex mountain topography, as demonstrated as an example in this paper, should assist in both the interpretation and explanation of contrasting paleoenvironmental conditions and for projecting future environmental conditions in alpine and/or high escarpment/mountain environments.

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