Late Pleistocene evolution of Scott Glacier, southern Transantarctic Mountains: implications for the Antarctic contribution to deglacial sea level

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Abstract

Glacial deposits preserved adjacent to Scott Glacier, southern Transantarctic Mountains, provide a record of past fluctuations in the thickness of the West Antarctic Ice Sheet. Geologic mapping of these deposits, in conjunction with emerging 10Be surface-exposure data, indicate that the most recent expansion of Scott Glacier occurred during the last glacial maximum in response to grounding of ice in the Ross Sea Embayment. At that time, the ice surface at the confluence of Scott Glacier and the West Antarctic Ice Sheet lay at ~1100 m elevation. While this ice-surface reconstruction is in accord with other geologic estimates from throughout the Ross Sea Embayment, it contrasts with most computer-based simulations, which tend to overestimate former ice thickness in the southern Ross Sea. Together with recently modelled estimates of Antarctica’s contribution to sea level, this finding calls into question an Antarctic source for meltwater pulse 1A.

1. Introduction

Most of the change in Antarctic ice volume since the last glacial maximum (LGM) occurred in West Antarctica, where grounded ice filled what are now marine embayments (e.g. Bentley and Anderson, 1998; Denton and Hughes, 2000; 2002; Anderson and Shipp, 2001). In the Ross Sea Embayment (RSE) (Fig. 1), for instance, the distribution of glacial landforms on the seafloor – including flutings, drumlins, till sheets, and the continuation of modern ice stream troughs – indicates that the ice sheet advanced as far as the outer continental shelf (Shipp et al., 1999; Anderson et al., 2002). This scenario is supported by geologic evidence from the Transantarctic Mountains (TAM), where outlet glaciers of the East Antarctic Ice Sheet (EAS) thickened in response to grounded ice in the Ross Sea (Mercer, 1968; Bockheim et al., 1989; Denton et al., 1989; Orombelli et al., 1990; Bromley, 2010), and from the McMurdo Sound region, where grounded Ross Sea ice dammed large proglacial lakes in the Dry Valleys (Hall et al., 2000). Though less well-documented, similar advances of the grounding line likely occurred in the Weddell, Amundsen, and Bellingshausen Seas and the Amery Basin (Elverhøi, 1981; Denton et al., 1992; Bentley and Anderson, 1998; Lowe and Anderson, 2002; Johnson et al., 2008; White et al., 2011).

Although the lateral extent of the WAIS at the LGM is relatively well defined, uncertainty remains over the former thickness of the ice sheet. Glacial-geologic evidence from the southern TAM suggests the ice sheet attained surface elevations of ≥1100 m at Reedy Glacier (Bromley et al., 2010), ~1250 m (Denton et al., 1989) or 1150 m (our observation: see footnote 1) at Beardmore Glacier, and 1100 m at Hatherton Glacier (Bockheim et al., 1989). Farther north, the LGM ice surface was ~710 m on eastern Ross Island (Denton and Marchant, 2000), ~640 m on Minna Bluff (Denton and Marchant, 2000), 350 m at Hjorth Hill (Hall and Denton, 2000), and ~400 m at Terra Nova Bay (Orombelli et al., 1990). Adjacent to the eastern RSE, Stone et al. (2003) reported geologic evidence from the Ford Range, Marie Byrd Land, suggesting the former ice-sheet surface there exceeded 1165 m elevation. Despite the consistency within the geologic dataset, these estimates differ from modelled reconstructions of the ice sheet, most of which infer thicker ice in the Ross Sea at the LGM. For example, the reconstructions of Denton and Hughes (2000, 2002), based on earlier glacial geologic data, call for ice-surface elevations of approximately 1600 m in the southern Ross Sea, whereas that of...

1 During a 2010 visit to Mt. Kyffin, Beardmore Glacier, we interpreted a drift edge on the peak’s northern flank as the LGM ice limit. Below the limit, white granite boulders are relatively unweathered, whereas above the limit boulders of the same lithology exhibit pronounced exfoliation and minor cavernous weathering. The drift edge lies at ~1150 m, approximately 100 m below an older yet more conspicuous drift limit.
Huybrechts (2002) prescribes even thicker ice (>2000 m surface elevation) along the southern TAM front. In contrast, estimates from Siple Dome (Fig. 1) are more in keeping with the new geologic evidence, suggesting relatively low surface elevations (~1000 m) close to the centre of the former ice sheet (Waddington et al., 2005; Price et al., 2007).

Deglaciation of the RSE following the LGM is resolved broadly on the basis of marine (Licht et al., 1996; Domack et al., 1999; Licht and Andrews, 2002; and references therein) and terrestrial radiocarbon data (Bockheim et al., 1989; Hall and Denton, 1999, 2000; Baroni and Hall, 2004), radar-based model data from Roosevelt Island (Conway et al., 1999), and surface-exposure ages from Reedy Glacier (Todd et al., 2010). These data indicate that (i) recession in the RSE occurred mainly during the Holocene and (ii) that grounding-line retreat may have slowed/stopped in recent millennia. In the north-east RSE, a comprehensive set of 10Be surface-exposure ages from the Ford Range indicates that post-LGM deglaciation in Marie Byrd Land was underway by at least 10 ka, with little change since 2–3 ka (Stone et al., 2003). A similar situation was reported by Todd et al. (2005) and Bentley et al. (2010) from the Marble Hills, eastern Ellsworth Mountains, and by Johnson et al. (2008) from the Pine Island Glacier region, where cosmogenic surface-exposure ages suggest a gradual and continued thinning of the WAIS until the Late Holocene. The former volume and deglacial history of the WAIS both are critical for understanding Antarctica’s contribution to global sea-level change during and since the LGM. This is particularly the case for hypotheses addressing the origins of deglacial events, such as meltwater pulse 1A (MWP-1A) (Fairbanks, 1989). Although it has been argued that this apparent ~20 m jump in sea level at 14.6 ka was caused by melting of northern-hemisphere ice sheets (Peltier, 1994; Flower et al., 2004), in particular the Laurentide Ice Sheet, modelling of post-glacial sea-level change suggests Antarctica as an alternative source (Clark et al., 1996, 2002, 2009; Weaver et al., 2003; Carlson, 2009). If true, corroborating evidence should be found in the glacial-geologic record from Antarctica, particularly in reconstructions of former ice-sheet volume and the timing of deglaciation. To begin to address this issue, we present a reconstruction of former ice-sheet thickness in the southern RSE based on a glacial-geologic record from Scott Glacier.

Scott Glacier (Fig. 1) is an outlet of the EAIS located in the southern TAM. Today, the glacier drains into the Ross Ice Shelf (Fig. 2). However, during the LGM and most of the subsequent deglaciation, Scott Glacier formed a tributary of the expanded ice sheet in the RSE. We used the glacial-geologic record from Scott Glacier to reconstruct the former surface elevation of this ice sheet at the LGM. Scott Glacier is ideal for this study for two reasons. First, the glacier is located in the far southern RSE and thus affords a close measure of LGM ice thickness from the heart of the former ice sheet. Second, glacial deposits are well preserved in the mountains alongside Scott Glacier, allowing for accurate reconstruction of the LGM ice-surface profile. These deposits form the basis of a cosmogenic 10Be chronology, which is the focus of a separate paper (Stone...
In addition to the late Pleistocene record, we describe older deposits at Scott Glacier. Occurring above the limit of relatively unweathered LGM deposits, weathered erratics, drift sheets, and moraines correspond to periods of earlier, more extensive glacier expansion and afford insight into the longer-term (Cenozoic) behaviour of glaciation in the southern TAM.

2. Surficial geology and geomorphology

Glacial deposits, such as moraines, drift sheets, and perched erratics, are abundant and well preserved along the margins of Scott Glacier, where katabatic winds and mountainous terrain maintain extensive areas of ice-free terrain. In the only previous glacial-geologic study, Mayewski (1975) assigned these deposits to three broad age groups, termed ‘lower drift’, ‘middle drift’, and ‘upper drift’, on the basis of relative weathering and position. Our mapping revealed a greater number of drift units, which are described in detail below. We focused on sites located along a ~200 km transect from the glacier mouth at the Karo Hills to its head at Mt. Howe (Fig. 2). At each site, we mapped deposits onto vertical aerial photographs and categorised these deposits on the basis of relative weathering, composition, morphology, and position relative to the modern glacier margin. Excavations enabled us to take clast and sediment samples from each unit so as to characterise glacial drifts on the basis of physical characteristics (e.g. weathering extent, composition). To reconstruct the former glacier surface profile, we measured elevations of the uppermost late Pleistocene deposits using a handheld GPS (estimated ± 15 m). For comparison, the modern glacier profile is based on 1:250,000-scale USGS topographic maps verified by our GPS measurements in key locations.

2.1. Karo Hills

At the glacier mouth, the Karo Hills (85°34′S, 154°00′W) form a small range of high-relief granite peaks overlooking the confluence of Scott Glacier and the Ross Ice Shelf (Fig. 2). The range is formed by Mts. Salisbury (952 m), Hastings (750 m), and Rigby (893 m) and is bounded to the west by Koerwitz Glacier, a tributary of Scott Glacier and currently the dominant source of glacial sediment for the site. Relatively warm conditions occur at the Karo Hills during summer, resulting in the formation of large supraglacial and ice-marginal meltwater ponds and streams. Glacially moulded, polished, and striated bedrock is widespread.

2.2. Taylor Ridge

Taylor Ridge (85°45′S, 154°00′W) is an alpine massif located 9 km up-glacier from Mt. Salisbury (Fig. 2) and dominated by the granitic peaks of Mts. Sletten, Nelson (1929 m), and Pulitzer (2156 m). Below Mts. Sletten and Nelson, a low-angled terrace (informally termed Nelson Platform) extends northward towards Koerwitz Glacier (Fig. 3). West of Nelson Platform is the broad Pulitzer Valley (informal name), which currently is occupied by a 6 km-long, north-flowing alpine glacier (Fig. 3). Other alpine glaciers on Taylor Ridge occur on the northern slope of Mt. Nelson and in the valley between Mts. Nelson and Sletten. Most deposits

![Fig. 3. Distribution of glacial deposits at Taylor Ridge and (inset) location map.](image)
described here are located on Nelson Platform and on the steeper slopes separating it from the modern ice margin.

2.3. Cox Peaks

South of Taylor Ridge, on the western margin of Scott Glacier, the long granitic spurs of Cox Peaks (86°02’S, 153°20’W) and the Walshe massif (86°18’S, 152°10’W) separate two major tributary glaciers: Vaughan and Bartlett. Cox Peaks comprises two parallel sub-ranges separated by a deep north-east-flowing valley, into which flows a lobe of Scott Glacier (Fig. 4). The terminus of this lobe is confluent with a local alpine glacier — Ragnarok Glacier (informal name) — in an area characterised by moraines and sizeable ice-marginal meltwater ponds. The topography of Cox Peaks is alpine, exhibiting horns, arêtes, and well-formed cirques (Fig. 5a). Small glaciers occupy several of the cirques.

2.4. Mt. Walshe

The Walshe massif is similar to Cox Peaks in that it consists of a deep valley hemmed in by alpine peaks and partially filled by a lobe of Scott Glacier (Fig. 6). Mt. Walshe (2050 m) itself is located on the precipitous eastern edge of the range, directly above the western margin of Scott Glacier (Fig. 6). As at Cox Peaks, the terminus of the Scott Glacier lobe is characterised by ice-cored moraines and an ice-covered meltwater pond.

2.5. LaGorce mountains: Ackerman Ridge and Surprise Spur

Ackerman Ridge (86°34’S, 148°05’W), located on the east side of Scott Glacier, forms the northern tip of the LaGorce Mountains and separates Scott Glacier from a major tributary, Robison Glacier (Figs. 2 and 7). In contrast to the higher peaks and plateaux in the range, the lithology of Ackerman Ridge is dominated by dark-coloured gneissic granite. Local topography is characterised by steep, high-relief summits separated by long arêtes, and valleys inundated by east-flowing lobes of Scott Glacier (Fig. 5b). Along the steep margins of each lobe, we observed meltwater ponds — some without an ice cover — and, on clear days, numerous meltwater streams. Though we mapped deposits throughout the northern section of Ackerman Ridge, preservation is best on the 4-km-long Surprise Spur, which protrudes from the highest point (2310 m) of the main crest (Fig. 7).

To complete the Scott Glacier transect, we visited three additional sites, including Mt. Gardiner (85°20’S, 151°00’W), Mt. Verlautz (86°46’S, 153°00’W), and Mt. Howe (87°21’S, 150°00’W) (Fig. 2). Although the glacial geology of these locations is not discussed in detail in this paper, each site provided elevation data for the upper limits of late-Pleistocene deposits and thus is included as part of the longitudinal-surface profile.

3. Late Pleistocene deposits at Scott Glacier

Along the length of Scott Glacier, a conspicuous unit of fresh drift with a clear upper limit is preserved on ice-free hillsides, commonly several hundred metres above the modern glacier surface. In places, the drift is thick and ice cored, whereas in others it comprises only a thin veneer of cobbles perched on the underlying surface. In all locations, however, the drift is loose and virtually unweathered. Mayewski (1975) classified this unit as part of the ‘middle drift’ and suggested it was deposited during the last glacial period. Since deposits of many different ages are preserved

![Fig. 4. Distribution of glacial deposits at Cox Peaks and (inset) location map.](image-url)
at Scott Glacier, we consider that terminology to be insufficient and here refer to this unit as Scott I drift. Because of its fresh appearance and stratigraphic position, as well as new surface-exposure data (Stone et al., in prep.), Scott I drift is correlative with the LGM and subsequent deglaciation.

3.1. Karo Hills

Abundant fresh erratics and ice-cored drift mantle the entire Karo Hills, indicating these summits were overridden by Scott Glacier during the most recent advance. Therefore, the site affords a minimum value of 952 m for the surface of the WAIS at the TAM front. Although there are numerous patches of stained, highly weathered bedrock in the Karo Hills, the occurrence of fresh, moulded, and striated bedrock, as well as unweathered clasts perched on oxidised bedrock, is suggestive of localised wet-based glaciation during the last glacial advance. This is in stark contrast to the Medina Hills, located less than 10 km away on the west side of Koerwitz Glacier, which, although surely covered by ice, exhibit rectilinear form and little ice-moulded bedrock. Cosmogenic surface-exposure ages of perched granite erratics collected from the north-west ridge of Mt. Rigby date to \( <12 \) ka and document the thinning of Scott Glacier from close to the LGM position to present (Stone et al., in prep.).

3.2. Taylor Ridge

Glacial deposits of many ages are preserved on the north-facing slopes of Taylor Ridge, overlooking the confluence of Scott and Koerwitz Glaciers (Fig. 3). During the last advance, ice from Scott Glacier flowed west across Taylor Ridge and deposited an extensive drift sheet bounded in places by a conspicuous moraine. At its highest point on the cliff overlooking Scott Glacier (Fig. 3), the Scott I drift edge is marked by a single low-relief (1 m high) boulder moraine at 1200 m elevation. This moraine is composed entirely of granite, in contrast to older, underlying deposits that include various exotic lithologies (e.g. dolerite). The elevation of this landform above Scott Glacier indicates the glacier surface was

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Fig. 5. (a) Alpine topography of Cox Peaks, with Scott Glacier visible in the background; (b) east-flowing lobe of Scott Glacier at Ackerman Ridge, LaGorce Mountains.

Fig. 6. Distribution of glacial deposits at Mt. Walshe and (inset) location map.
~600 m higher during the Scott I advance than it is today (Table 1). No unweathered (i.e. Scott I) erratics occur on the slopes above the moraine.

On the broad Nelson Platform and in Pulitzer Valley the drift edge is marked by a closely spaced pair of moraine ridges (Fig. 3), as much as 1–3 m in relief, composed of loose sandy gravel, cobbles, and boulders. The elevation of these moraines descends from 1090 m to 1050 m from east to west, indicating the former confluence of Scott and Koerwitz Glaciers was farther west than today. Elsewhere on Taylor Ridge, Scott I drift is ice cored and exhibits widespread frost cracking. In contrast to other sites visited, the abundance of weathered material in Scott I drift at Taylor Ridge is relatively high. For example, granite boulders exhibiting light staining and exfoliation are common and in many cases are perched upon boulders of much fresher appearance, suggesting transport of previously weathered material.

3.3. Cox Peaks

The Scott I drift edge at Cox Peaks typically is represented by the upper limit of fresh erratics on bedrock. An exception occurs above Ragnarok Glacier, where the drift limit is marked by a conspicuous metre-high lateral moraine extending for almost 2 km. The limit slopes westward towards the high peaks of the Faulkner Escarpment, from 1390 m above the trunk of Scott Glacier to 1230 m where the Scott lobe intersects Ragnarok Glacier (Fig. 4). This slope indicates that the former confluence of Scott Glacier and the alpine glacier lay at least 7 km west of its present position. At this time, the surface of Scott Glacier at Cox Peaks was approximately 400 m higher than today.

Scott I drift at Cox Peaks is characterised by extensive deposits. However, because only the summits and higher crests protruded above the ice, well-defined moraines are rare and much of the former ice surface is defined by the upper limit of fresh, unweathered boulders. One exception is on the steep colluvial slopes above the southern edge of Ragnarok Glacier. Here, the expanded Scott Glacier deposited a thick drift of white granite boulders, cobbles, and gravel, the upper limit of which forms a conspicuous edge of ~1 m relief.

Drift thickness and composition varies with elevation at Cox Peaks, so that Scott I deposits in the valley bottoms are thicker than those on the peaks themselves. Higher up, the till comprises coarse gravel and small (<1 m relief), unweathered erratics perched upon stained and grussied bedrock and patches of older, highly weathered till (Fig. 8a,b). Erratics commonly exhibit moulding. At lower elevations, Scott I drift typically is much thicker, is characterised by large (1−2 m relief), rounded boulders, and contains abundant silt. Whether this silt, which is indicative of wet-based glacial erosion, was produced locally or reflects erosion and incorporation of older deposits is unclear. These thicker deposits commonly contain buried ice that, on days of high solar irradiance, produces small meltwater-fed debris flows.

Since the last glacial high-stand, thinning of Scott Glacier and/or advance of local ice has caused the Scott-Ragnarok confluence to migrate down-valley to its current position. Throughout this terminus area, desiccated algal remains attest to the existence of ice-dammed lakes. The most recent landform deposited by the Scott lobe is a 2−3 m relief, ice-cored lateral-terminal moraine bounding the ice tongue on its northern and western margins and abutting the terminal moraine of Ragnarok Glacier (Fig. 4).

### Table 1

LGM ice-surface elevations at sites along Scott Glacier.

<table>
<thead>
<tr>
<th>Site</th>
<th>Approximate distance from modern grounding line (km)</th>
<th>LGM surface elevation (m)</th>
<th>ΔLGM-modern surface change (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Karo Hills</td>
<td>0</td>
<td>&gt;950</td>
<td>&gt;750</td>
</tr>
<tr>
<td>Taylor Ridge</td>
<td>8</td>
<td>1200</td>
<td>600</td>
</tr>
<tr>
<td>Cox’s Peaks</td>
<td>28</td>
<td>1375</td>
<td>375</td>
</tr>
<tr>
<td>Mt. Walsh</td>
<td>40</td>
<td>1600</td>
<td>400</td>
</tr>
<tr>
<td>Mt. Gardiner</td>
<td>55</td>
<td>1650</td>
<td>400</td>
</tr>
<tr>
<td>LaGorce Mtns.</td>
<td>75</td>
<td>2100</td>
<td>400</td>
</tr>
<tr>
<td>Mt. Verlautz</td>
<td>79</td>
<td>2200</td>
<td>200</td>
</tr>
<tr>
<td>Mt. Howe</td>
<td>120</td>
<td>2450</td>
<td>&lt;20</td>
</tr>
</tbody>
</table>

Fig. 7. Distribution of glacial deposits at LaGorce Mountains and (inset) location map.
contrast with most late Pleistocene deposits elsewhere along Scott Glacier, this moraine contains abundant silt in addition to gravel and boulders.

3.4. Mt. Walshe

During the last expansion of Scott Glacier, only the summits of the Mt. Walshe massif protruded above the ice surface. At that time, the west-flowing lobe of Scott Glacier, which today terminates at ~1000 m, advanced into the valleys and cirques, reaching elevations of between 1240 and 1300 m. This ice coalesced both with the enlarged Souchez Glacier (a tributary of Scott Glacier) and with local alpine glaciers (Fig. 6). Scott I deposits at Mt. Walshe largely comprise thin, granite-dominated drift overlying older, weathered deposits and bedrock. In general, the Scott I drift edge is relatively indistinct and marked by the upper limit of occasional fresh, unweathered erratics. One exception is at the mouth of Achilles Valley, where the unit is bounded by a conspicuous boulder moraine as much as 3 m high and composed of fresh pink granite. Similarly, on the north-east shoulder of Mt. Walshe the Scott I drift edge forms a series of discontinuous ridges of unweathered granite boulders, cobbles, and gravel — remnants of former ice-cored moraines. These landforms are located directly above the western margin of Scott Glacier (local surface elevation ~1200 m) at an elevation of ~1600 m, indicating that the magnitude of glacier thickening during deposition of Scott I drift in the vicinity of Mt. Walshe was on the order of 400 m relative to today (Table 1).

Several prominent moraine systems deposited by the retreating Scott Glacier lobe are well-preserved in the broad valley sloping down towards the Scott lobe immediately west of Mt. Walshe (Figs. 6 and 9). The highest occurs at 1180 m below the mouth of Achilles Valley and is a 2 m-high, flat-topped moraine of white granite gravel and cobbles. A second moraine corresponding to this same limit is preserved at ~1150 m elevation below the mouth of Methuselah Valley. Farther down-valley, and closer to the Scott lobe, a prominent moraine ridge (2–3 m relief) composed predominantly of gneissic Mt. Gardiner granite crosses the valley. This ridge, which consists of stacked boulders and cobbles and is in places ice cored, corresponds to a former terminus elevation of 1095 m. Below this limit, between 1035 and 1055 m elevation, the valley floor is relatively flat and dominated by frost-cracked pavements of compacted silt, sand, and gravel with abundant desiccated algae. Further investigations here revealed abundant desiccated algae. The youngest recessional limit forms a horizontal terrace of gravel and small cobbles at 1015 m elevation. Algae and compacted silty sand occur on the slope separating this moraine from the modern glacier margin.

Fig. 8. Unweathered erratics of Scott I age: (a) perched clast on weathered bedrock, Cox Peaks; (b) boulder on weathered till, Cox Peaks; (c) light-coloured granite erratic deposited by Scott Glacier, LaGorce Mtns.; (d) dark-coloured metasedimentary erratic deposited by Robison Glacier, LaGorce Mtns.

Fig. 9. View west from Mt. Walshe, showing the current margin of the Scott lobe (bottom right) and the distant Faulkner Escarpment. The recessional moraines on the valley floor correspond to retreat of the Scott Lobe following the LGM.
3.5. LaGorce mountains: Ackerman Ridge and Surprise Spur

Scott I deposits on Ackerman Ridge typically comprise a thin drift of unweathered boulders and cobbles perched on bedrock and talus at elevations up to 2100 m. There is a strong lithological difference between material transported by Scott Glacier and that deposited by a major tributary, Robison Glacier, which bounds Ackerman Ridge to the northeast. Whereas the former is dominated by pale granite (Fig. 8c), Robison ice deposited predominantly dark-coloured metasedimentary clasts (Fig. 8d). Consequently, the distribution of granitic and metasedimentary erratics provides an indication of former ice configuration at Ackerman Ridge.

Unweathered metasedimentary erratics on the ridge crest and a conspicuous drift limit west of the 1800 m col (Fig. 7) indicate that west-flowing ice from the thickened Robison Glacier crossed Ackerman Ridge during the last advance. Meanwhile, a NE-flowing tongue of Scott Glacier deposited a thin drift of granite boulders on the slopes of Surprise Spur. The confluence of these two ice lobes occurred at 1850 m elevation, approximately one hundred metres west of the col, as suggested by the convergence of metasedimentary and granitic drift. At that time, the ice surface was ~200 m higher than present adjacent to Surprise Spur and up to 300 m higher in the vicinity of the main Ackerman Ridge.

Thick accumulations of till are preserved on the lower slopes of LaGorce Mountains, such as at the end of Surprise Spur. There, Scott I drift is ice-cored and exhibits extensive frost cracking. This unit contains abundant rounded and moulded boulders, in addition to large quantities of silt, indicative of wet-based glaciation. As at Cox Peaks, it is not known whether these conditions exist in the vicinity of LaGorce Mountains or farther afield — beneath the EAI, for instance — or whether they represent reworked deposits from older glaciations.

Deglaciation at LaGorce Mountains was interrupted by several episodes of margin stability or advance, as evidenced by a series of low (~1 m relief), discontinuous moraine ridges on the slopes above the ponds (Fig. 7). A more prominent limit, known informally as the Amazing Moraine, occurs throughout Ackerman Ridge and Surprise Spur, forming a boudery ridge 20–30 m above the glacier edge (Fig. 10a). This conspicuous landform, which in places is ice cored, is up to 1.5 m in relief and is composed primarily of stacked, unweathered granite boulders and coarse gravel. The boulders are predominantly sub-angular in shape, though some are moulded and striated.

4. Pre-Scott I deposits at Scott Glacier

Deposits corresponding to earlier, pre-Scott I advances are preserved in several locations alongside Scott Glacier. Although the fragmentary nature of this record precludes meaningful correlation among sites, these older deposits indicate that the glacier underwent numerous periods of expansion prior to the most recent, Late Pleistocene event. Here, we describe the distribution and morphology of older deposits at Taylor Ridge, Mt. Walshe, and LaGorce Mountains.

4.1. Taylor Ridge

The most complete glacial record is from Taylor Ridge, where at least nine separate units are preserved on Nelson Platform (Fig. 3). The youngest of these deposits — unit TR2 — lies immediately upslope of the Scott I limit and, with the exception of more extensive weathering characteristics, is similar in composition, morphology, and orientation to Scott I drift. This unit (TR2) comprises a deflated, frost-cracked drift of granitic boulders, most of which exhibit exfoliation and light staining, with minor amounts of desert varnish. TR2 extends only a short distance (~50 m horizontally, ~15 m vertically) beyond the Scott I limit and, at its outermost extent, forms a thin veneer of drift overlaying units TR3–7. An exception is in Pulitzer Valley, where the drift edge is marked by a metre-high rampart (Fig. 3).

Above the TR2 limit, five closely nested lobes of weathered till extend to 1190 m on Nelson Platform. These deposits (TR3–7 in Fig. 3) vary only slightly in weathering from youngest to oldest. Buried ice is widespread throughout, though it occurs at greater depth (~60 cm in the outer lobes than in the inner units (~30–40 cm). The outermost lobe, TR7, is bounded by a moraine (1–2 m relief) composed of perched granite boulders. These lightly stained and heavily exfoliated clasts exhibit pits of 1 cm depth and minor cavernous weathering. We traced the moraine west into Pulitzer Valley, where it is overrun by the alpine glacier (Fig. 3). Each of the nested lobes exhibits a similar moraine on the broad ridge separating Nelson Platform and Pulitzer Valley. Here, the moraines are as much as 3 m in relief. Farther west and east, however, the moraines disappear.

The closely spaced nature of the lobe drifts (TR3–7), in addition to the relatively minor weathering differences from oldest to youngest, suggests they were formed when a large area of stranded supraglacial moraine was let down onto Taylor Ridge during glacier thinning. This inference is supported by the distribution of the deposits in Pulitzer Valley, where the pattern of individual lobes, so apparent on the platform, dissolves into a single drift sheet. The surface of this drift sheet becomes gradually more weathered with increasing distance from the glacier, as on Nelson Platform, and the general weathering characteristics of the unit also are the same as on Nelson Platform. A modern analogue for the formation of the TR3–7 lobes exists at Mt. Howe, where thick deposits of supraglacial

Fig. 10. (a) Oblique aerial view of the ‘Amazing moraine’ at Ackerman Ridge, LaGorce Mtns.; (b) Scott I drift edge on Surprise spur, LaGorce Mtns., marked by the line of light-coloured, unweathered granite boulders. Upslope (left) of the drift limit, boulders of Scott 2 age exhibit pronounced staining as well as minor pitting and exfoliation.
moraines have accumulated in the lee of the nunatak. These accumulations exhibit individual lobes and ridges and are noticeably more weathered with increasing distance from the 'clean' glacier surface.

Above and distal to the TR7 lobe, at least three units of severely weathered till are preserved on the lower slopes of Mt. Sletten, Nelson, and Pulitzer. The youngest deposit (TR8; Fig. 3) is a thin drift of predominantly granite boulders extending up to 1325 m elevation. Boulders associated with this unit are severely exfoliated and stained, and exhibit moderate (up to ~30 cm diameter) cavernous weathering. TR9 is preserved as patches on Mt. Nelson and Mt. Pulitzer, where it forms a drift of boulders, some as large as 4 m, perched on the underlying surface to at least 1350 m elevation (Fig. 3). The boulders, many of which have been split by frost action, exhibit severe staining and caverns of 0.5–1 m diameter. The oldest identified deposit on Taylor Ridge, unit TR10, underlies both TR8 and TR9. This silt-rich drift typically forms a deflated surface of cobbles and gravel, in which are set the rotten cores of granite and dolerite boulders, some with salt crusts of 2-cm thickness. Moraines (4 m relief) are a common feature of TR10 and form series of large ridges to an elevation of ~1390 m on the slopes of Mt. Nelson (Fig. 3). In size, composition, and position these landforms are similar to the Reedy E moraines in the Quartz Hills, Reedy Glacier, dated to at least 5 Ma (Bromley et al. 2010). Elsewhere on Taylor Ridge, the upper limit of TR10 has been removed by slope processes or obscured by deposits from three alpine glaciers.

4.2. Mt. Walshe

Pre-Scott I deposits are widespread throughout the Mt. Walshe area. However, they typically occur only as patches of weathered boulders and sediments, visible where overlying deposits are thin. A notable exception is at ~1600 m on the narrow north-east shoulder of Mt. Walshe (Fig. 6), where deposits of many ages (including Scott I) overlie a deflated, severely weathered surface of silt-rich gravel. This basal unit (MW2), which overlies bedrock, contains abundant clasts of pale-coloured granite sourced from Mt. Gardiner, most of which are rotten and salt-encrusted to a depth of 50 cm. Small (~50 cm relief), perched granite boulders in various states of weathering, ranging from minor staining and exfoliation to almost complete disintegration, are embedded in and perched upon MW2 drift. Although several episodes of glaciation are recorded, it not possible to separate the units by age, nor to identify corresponding drift limits, due to their varying nature and limited spatial exposure.

Higher on Mt. Walshe, pre-Scott I deposits are defined with greater clarity. For instance, at 1670 m, weathered granite bedrock gives way to a black dolerite pavement overlying a silt-rich, matrix-supported diamicton (MW4 drift; Fig. 6). This 1–2 m thick unit contains equal parts granite and dolerite clasts, the former being ghosts to a depth of 50 cm. Dolerite lasts are unweathered throughout. Below 50 cm, clasts are well preserved regardless of lithology. Indeed, the pink granite bedrock underlying the unit is freshly moulded and appears to have undergone no weathering whatsoever. Only minor quantities of granite clasts are present on the surface, most of which are small (~30 cm) and rotten. Salt accumulations occur beneath most surface clasts. This silt unit has been cut on the north side by subsequent valley erosion, exposing the deposit in section. On the surface, the exposure consists of grey, weathered silt with in-situ boulders and cobbles. However, removal of this weathered surface revealed 25 cm of finely laminated silts that, subsequent to deposition, have been broken into fragments 1–5 cm in length to form a breccia. Below 25 cm depth, the fragments are cemented into buried ice. With the exception of large fragments that appear to have rolled down the face of the exposure, there are no clasts associated with these water-lain, brecciated sediments.

MW4 drift extends horizontally for approximately 100 m and reaches an elevation of 1700 m, where it is overlain by a sandy, clast-supported diamicton (MW3 drift). This younger unit, which does not occur at lower elevation, extends to the base of the summit cliff at 1780 m. It is a loose, unconsolidated drift with widespread buried glacier. The surface of the younger unit is characterised by white granite boulders, all of which are moderately exfoliated, stained, and cavernously weathered.

4.3. LaGorce Mountains

Patches of weathered drift corresponding to at least three glacial episodes are preserved above the Scott I limit on Surprise Spur (Fig. 6). Above 1850 m, the youngest pre-Scott I unit, LG2, comprises a thin, patchy drift of predominantly granite boulders overlying a much older (LG4) till surface. The LG2 boulders, though heavily stained, are only slightly exfoliated and pitted (Fig. 10b). Some basalt cobbles exhibit wind polish. The indistinct upper limit of LG2 drift occurs at ~1870 m. A second, more-weathered drift unit (LG3) extends to approximately 1950 m elevation. Again, this deposit forms a patchy veneer of boulders, cobbles, and gravel overlying a significantly older (LG4) surface. Boulders are visibly more weathered than those in LG2 drift and display deep (~5 cm) pits, extensive exfoliation and staining, and, in some cases, desert varnish.

The deflated, highly weathered LG4 till upon which LG2 and LG3 are draped extends to a conspicuous limit at 2025 m beneath the crest of Ackerman Ridge (Fig. 7). This deposit displays a well-sorted surface of gravel, weathering detritus, and shattered blocks, into which are set cobbles and small boulders of many different lithologies. Below the surface, silty sand is abundant. Boulders associated with LG4 are cavernously weathered and rotten, and exhibit severe staining and exfoliation. In places where this drift is thin, an even older, highly compacted drift surface is visible, though its morphology and extent are unknown.

5. Former surface profile of Scott Glacier

The upper limit of fresh, unweathered deposits on hillsides above Scott Glacier, commonly marked by a moraine or drift edge, forms the basis for reconstructing the longitudinal ice-surface profile for the Scott I glaciation, shown in Fig. 11. This profile is based on measurements from each site, including Mt. Gardiner and Mt. Verlautz (Fig. 2). At the glacier mouth, where no upper limits of Scott I drift occur because the mountains were overrun, the profile is extrapolated to provide minimum, maximum, and likely reconstructions. Based on the occurrence of Scott I deposits on the summit of Mt. Rigby, the minimum reconstruction places the former ice surface at ~1000 m and results in a relatively steep glacier profile. Our maximum reconstruction assumes a horizontal ice surface down-glacier of Taylor Ridge, the closest site to the glacier mouth with good elevation control (1200 m). There is no evidence to indicate that the ice-surface elevation exceeded our maximum reconstruction. If it had, flow of Scott Glacier would have reversed.

Our preferred profile was constructed by extrapolating down-valley the surface profile between Taylor Ridge and Mt. Walshe, and suggests the former surface elevation of Scott Glacier was approximately 1100 m at the confluence with the WAIS at the TAM front (Fig. 11). We note, however, that the profile may be time-transgressive in that maximum ice thickness could have occurred earlier at the mouth than at the head of the glacier. Such a pattern was identified by Todd et al. (2010) at Reedy Glacier and was
attributed to the concurrent effects of increased precipitation over the EAIS and retreat of the WAIS during the Holocene. If a similar situation occurred at Scott Glacier, the former surface profile would have been somewhat shallower than our preferred profile. Additionally, our reconstructions do not account for isostatic rebound of the TAM, due to either glacial downcutting or changes in ice-sheet mass since the LGM, because any vertical displacement is unlikely to have exceeded the GPS uncertainty (±15 m).

The pattern of thickening of Scott Glacier during the Scott I glaciation (Fig. 11) is similar to that observed for other TAM outlet glaciers (e.g. Mercer, 1968; Bockheim et al., 1989; Denton et al., 1989; Orombelli et al., 1990; Denton and Hall, 2000; Bromley et al., 2010; Todd et al., 2010) that are correlated with the LGM (Denton and Hughes, 2000). At neighbouring Reedy Glacier, for example, the age of these youngest deposits (Reedy III drift; Mercier, 1968; Bromley et al., 2010) at their upper limits ranges from 17 ka near the glacier mouth to 7–8 ka near the East Antarctic plateau (Todd et al., 2010). Farther north, in the Dry Valleys region, correlative Ross Sea drift yielded maximum ages of ~14–17 ka (Hall and Denton, 2000). At Scott Glacier, surface-exposure ages from the Scott I drift also indicate an LGM age for the deposits (Stone et al., in prep.). On these bases, we are confident that Scott I drift represents thickening of the glacier during the LGM.

Unlike at neighbouring Reedy Glacier, the highly fragmentary distribution of pre-LGM deposits at Scott Glacier precludes accurate correlation between sites. Nonetheless, deposits at Scott Glacier become progressively older with increasing elevation above the ice surface, a pattern also observed at Reedy Glacier (Mercier, 1968; Bromley et al., 2010), Beadmore Glacier (Denton et al., 1989), and Hatherton Glacier (Bockheim et al., 1989). While at face value this pattern might suggest that earlier expansions of Scott Glacier were more extensive than later episodes, in keeping with global patterns (Mercer, 1976), Bromley et al. (2010) proposed that a similar pattern at Reedy Glacier may be the combined product of glacial downcutting and progressive shrinkage of the WAIS. The latter process was attributed to enhanced drainage efficiency of the ice sheet in response to widening and straightening of subglacial troughs (e.g. Evans, 1969), with the result that successive expansions of the WAIS would become smaller. We suggest that similar processes produced the long-term lowering of Scott Glacier.

Both the composition and morphology of Scott I drift are characteristic of a cold-based ablation till. Drift typically consists of angular clasts, sand, and gravel, is loose and unconsolidated in nature, and contains widespread buried ice. A notable exception occurs in the LaGorce Mountains, where abundant silt and rounded boulders on the lower slopes of Ackerman Ridge and Surprise Spur are indicative of wet-based glaciation. Considering the cold prevailing climate at Scott Glacier, it is unlikely that basal melting in the vicinity of LaGorce Mountains is the source of this material. A more plausible scenario is that freezing-on of basal material is occurring in high-pressure regions, such as beneath the thickest parts of Scott Glacier or where the EAIS abuts the TAM at the glacier head (T. Hughes, personal communication). An alternative explanation is that the deposits represent reworking of sediments of the much older Sirius Formation, which are characterised by abundant silt and glacially moulded clasts (Denton et al., 1993).

Fig. 11 shows the minimum, maximum, and preferred surface profiles of Scott Glacier during the Scott I (LGM) advance relative to today, based on the upper limits of fresh drift. These reconstructions indicate that the former ice surface must have been between ~1000 and 1200 m elevation at the mouth of Scott Glacier, while our preferred profile suggests an elevation of ~1100 m. This reconstruction is in accord with geologic evidence from other EAIS outlet glaciers (e.g. Bockheim et al., 1989; Denton et al., 1989), including Reedy Glacier at the head of Mercer Ice Stream (Bromley et al., 2010; Todd et al., 2010) and from Marie Byrd Land (Stone et al., 2003) (Fig. 1). Reconstructions based on ice-core data from Siple Dome in the central RSE suggest a somewhat lower ice-sheet surface (800–1000 m; Waddington et al., 2005; Price et al., 2007). As discussed by Bromley et al. (2010), the geologic and Siple Dome datasets might be reconciled if thicker ice occurred along the front of the TAM and in Marie Byrd Land and thinner ice occupied the central RSE. This scenario implies that the WAIS ice streams remained active throughout the LGM, as suggested by Parizek and Alley (2004).
Relative to our reconstructions, previous determinations of former ice thickness in the southern RSE are larger. For example, Denton and Hughes (2002) placed the LGM ice surface near the mouth of Scott Glacier ~500 m above our reconstruction, likely because their reconstruction was based on moraines at Reedy Glacier that were undated at the time. An even greater difference (up to 1000 m) exists between our estimate and those of Stuiver et al. (1981), Nakada and Lambeck (1988), Huybrechts (1990, 2002), and Ritz et al. (2001), all of which called for an ice-surface elevation of ~2000 m in the vicinity of Scott Glacier. More recently, several models have produced ice-surface elevations that overlap broadly with our reconstruction (Pollard and DeConto, 2009: 1000–2000 m; Mackintosh et al., 2011: 1000–2000 m; Whitehouse et al., 2012 1000–1500 m). In contrast, Ivins and James (2005), incorporating several geologic and ice-core data points in West Antarctica (e.g. Ackert et al., 1999; Cowdery et al., 2004), estimated ice elevation in the southern RSE of 600–700 m.

The general inconsistency among reconstructions is not restricted to the southern RSE. Based on the distribution and chronology of glacial deposits throughout Antarctica (e.g. Ackert et al., 2007; Mackintosh et al., 2007; Bentley et al., 2010; Hein et al., 2011), it appears most model simulations have overestimated LGM ice volume. While this discrepancy emphasises the persistent challenge of conforming glaciologic models to field data, it also calls into question existing estimates of Antarctica’s contribution to sea-level change following the LGM and is particularly pertinent to the debate over the likely source of MWP-1A.

If, as suggested by Clark et al. (1996) and Bassett et al. (2005), Antarctica was the principal source of this event, a volume of ice equivalent to ~20 m sea-level rise would have been removed from the Antarctic ice sheets at ~14.6 Ka. Moreover, this collapse would have taken place in 500 years or less (Fairbanks, 1989; Hanebuth et al., 2000). However, with the exception of earlier models, which do not conform to updated field data (e.g. Stuiver et al., 1981; Nakada and Lambeck, 1988), and despite significant variability among simulations, more recent modelled reconstructions converge in showing that there was insufficient ice in Antarctica during the LGM to account for MWP-1A (Table 2). As pointed out by Bentley (1999) and, more recently, Mackintosh et al. (2011), the total volume of excess ice (relative to today) in Antarctica was closer to 10 m than to 20 m. Furthermore, since these models tend to overestimate former ice-sheet volume, the overall contribution of Antarctica to post-LGM eustatic sea-level rise probably was less, making an Antarctic source for any large-scale jump in sea level even less viable.

In addition to this fundamental problem with former ice-sheet volume, the deglacial chronology also is incompatible with an Antarctic source for MWP-1A. Consisting of radiocarbon (e.g. Bockheim et al., 1989; Licht et al., 1996; Hall and Denton, 2000; Licht, 2004), surface-exposure (Stone et al., 2003; Ackert et al., 2007; Mackintosh et al., 2007, 2011; Todd et al., 2010), and ice-core data (Price et al., 2007) from sites in both West and East Antarctica, this chronologic record shows consistently that the onset of deglaciation was, at the earliest, coincident with MWP-1A but most likely post-dated the event. Indeed, Ackert et al. (1999) showed from Mt. Waeschle that the central WAIS did not thin from its maximum position until ~10 Ka. Moreover, as shown in the Ross (Hall and Denton, 2000; Stone et al. 2003) and Weddell Sea Embayments (Bentley et al., 2010), in Mac. Robertson Land (Mackintosh et al., 2011), and in the circum-Antarctic marine record (e.g. Seigert et al., 2008; and references therein), rates of ice-sheet retreat initially were low and did not increase until the Holocene, several millennia after MWP-1A.

In contrast to this growing consensus, a recent interpretation of marine sediments in the Weddell Sea invokes an earlier onset for retreat (Weber et al., 2011). Citing the transition from varved to bioturbated sediments as evidence for ice-sheet recession, Weber et al. suggested that deglaciation of the EAIS in the eastern Weddell Sea began ~19 Ka, several millennia before the onset of ice thinning in the Ellsworth Mountains (Bentley et al., 2010). If true, the pattern of Antarctic deglaciation might have exhibited more localised variability, perhaps reflecting factors such as bathymetry; nonetheless, the possibility of such small-scale variability does not diminish the robust evidence establishing the relatively late deglaciation of Antarctica following the LGM.

7. Conclusions

The distribution and composition of glacial deposits along the length of Scott Glacier show that the youngest unit — Scott I drift — was deposited during the LGM and therefore is coeval with Reedy III drift at Reedy Glacier, Beardoie drift at Beardmore Glacier, Britannia drift at Hatherton Glacier, Ross Sea drift in McMurdo Sound, and Terra Nova drift at Reeves Glacier.

Grounding of the WAIS in the Ross Sea during the LGM caused the lower region of Scott Glacier to thicken significantly. Our reconstruction of the former glacier profile, based on the upper limits of Scott I drift, indicates that the LGM ice surface elevation was approximately 1100 m at the confluence of Scott Glacier and the WAIS. This estimate is consistent with geologic–based reconstructions from throughout the RSE but conflicts with previous reconstructions that invoke thicker Ross Sea ice during the LGM. An implication is that there likely was insufficient excess ice in Antarctica to account for MWP-1A.

The distribution and sedimentology of pre-LGM deposits indicates that Scott Glacier has undergone numerous periods of expansion since the inception of polar conditions. Prior to that time, widespread wet-based glaciation is suggested by MW4 drift and by the relict alpine topography. Based on relative weathering, deposits at Scott Glacier generally increase in age with increasing elevation. As at Reedy Glacier, this pattern likely reflects the combined influences of glacial downcutting and progressive shrinkage of the WAIS.

Acknowledgements

This project was supported by the Office of Polar Programs of the National Science Foundation (Grant number ANT-0636687). We are grateful for the excellent logistical support provided by Ken Borek Air, the Air National Guard, and Raytheon Polar Services. Also, we are indebted to Maurice Conway, Seth Cowdery, and Nathan Mietkiewicz for invaluable assistance in the field, to Terry Hughes for technical discussions, and to two anonymous reviews for providing thoughtful, in-depth, and constructive comments.


