Utilisation of remote sensing for the study of debris-covered glaciers: development and testing of techniques on Miage Glacier, Italian Alps

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CHAPTER 1: INTRODUCTION

1.1. Background

Mountain glaciers are among the best natural indicators of global climate change (Paul et al., 2003; Bolch and Kamp, 2005; Haeberli et al., 2007), with any changes in their physical properties (such as size, mass balance, debris cover and velocity) occurring as a response to changes in air temperatures and precipitation (Ranzi et al., 2004). One of the most serious consequences of global warming is glacier recession (Haeberli et al., 2007). European alpine glaciers lost around half of their total volume between 1850 and 1975, and a further 25% between 1975 and 2000 (Paul et al., 2007; Haeberli et al., 2007).

The response of alpine glaciers to climate warming has also been identified as a key factor in melt water runoff variations and resulting sea level changes (Diolaiuti et al., 2003), with the IPCC (Intergovernmental Panel on Climate Change), highlighting that melt from mountain glaciers and ice caps alone (excluding ice sheet, ice shelves, and sea ice melt) contributed to 0.77 +/- 0.22 mm year of sea level rise between 1991 and 2004 (Lemke et al., 2007). Furthermore, there are localised impacts upon the geomorphology and hydrology of the valleys that these glaciers occupy (Stokes et al., 2007). The monitoring of these mountain glaciers and their dynamics, therefore, provides an important contribution to climate monitoring (Kaab, 2002; Haeberli et al., 2007; Zemp et al., 2009).

In addition to sea level rise and impacts on lowland water resources, another key consequence of glacial retreat is mountain slope instability, which occurs as glaciers retreat upvalley or if surface lowering is experienced. Consequently, the glacial ice no longer provides support for the steep valley sides (debuttressing) (Deline, 2009). When combined with increased permafrost melting due to increasing air temperatures and changes in insulation provided by snow cover (a
result of global warming) (Lemke et al., 2007), this increases the instability of the valley sides, resulting in higher rates of rockfall supply to a glacier (Stokes et al., 2007; Deline, 2009; Diolaiuti et al., 2009). This increase in debris input combined with reduced glacier transport rates (Kirkbride, in press) (itself a result of a decrease in ice inputs into the glacier due to climatic warming), and melt out of pre-existing englacial sediment (Diolaiuti et al., 2009), often results in a glaciers surface becoming debris-covered.

Once debris material covers the full width or at least part of the ablation zone the glacier can be defined as debris-covered (Nakawo et al., 1993; Mattson et al., 1993; Singh et al., 2000). Others state that approximately 50% of a glaciers ablation zone must be covered in debris material to be classified as debris-covered (Kirkbride, in press). At present, debris-covered glaciers are widespread in the mountain regions of Asia, the Himalayas, New Zealand, and Alaska, with a smaller number found in Europe (Smiraglia, et al., 2000; Kellerer-Pirklbauer et al., 2008; Diolaiuti et al., 2009).

The need for increased monitoring is, therefore, justified by the likely increase in debris cover on glaciers in the future (through the influence of glacial retreat and permafrost melting), resulting in the development of a debris-covered glacier (Kellerer-Pirklbauer et al., 2008). As a consequence, future glacial monitoring will have to consider a supraglacial debris cover as a common occurrence on a glacier surface (Paul et al., 2003). However, currently debris-covered glaciers are under-studied. Therefore, debris-covered glaciers require special attention and increased monitoring due to the impact of a debris layer on patterns of surface ablation and thinning, along with the subdued response of debris-covered glaciers to climate changes, which is significantly different to a clean glaciers response of glacial retreat.
Consequently, when glaciers become covered in debris their dynamics and geomorphological processes are affected significantly (Deline, 2005) due to the influence of debris cover on sub-debris ice melt and mass balance (Kirkbride, *in press*). The presence of a debris layer will, therefore, also have an impact upon melt rates from a glacier, with the preservation of ice and freshwater supplies (by a thick debris layer) for a longer period than predicted by melt models for a ‘clean’ glacier. Subsequently, this has an impact upon predicted run-off rates from debris-covered glaciers, which are a key freshwater resource. Also of significance, is that the impact of debris layers upon glacial surface melt and their retreat has not been included in model calculations of predicted global sea level rise due to glacial melt and retreat over the next century (Church *et al.*, 2001).

The importance of present research in increasing the understanding of debris-covered glaciers and their characteristics is therefore highlighted, along with monitoring the impacts of an increasingly warmer climate on predicted run off rates. Any variations in run off rates require monitoring as they will affect both those dependant on the glacier as a freshwater resource and rates of predicted sea level rise. In turn this demonstrates the requirement for the inclusion and increased monitoring of debris-covered glaciers within current global glacier monitoring programs. The full impact of a debris layer and the key role of debris thickness is outlined below.

The effect of debris cover on a glacier is varied and depends significantly upon the debris thickness and its thermal properties as this influences the heat transfer through the debris cover into the ice below. This means debris can have either a positive or negative impact upon surface ablation (Ostrem, 1959; Nakawo and Young, 1981; Pelto, 2000; Kellerer-Pirklbauer *et al.*, 2008; Diolaiuti *et al.*, 2009; Kirkbride, *in press*). Thin debris and a dispersed debris layer will enhance ablation, because the lower-albedo debris material will absorb more of the
incoming shortwave radiation compared to the higher albedo bare ice. The resulting increase in debris temperature means that the heat is conducted through it to the ice below, resulting in greater levels of ablation.

In a thicker debris layer, however, ablation will be reduced due to the insulating effect of the thick debris layer, with incoming short-wave energy absorbed by the debris layer or re-radiated back to the atmosphere as longwave radiation. Therefore, the energy is used to heat up the debris layer itself or is lost back to the atmosphere and is not all transported to heat/melt the ice below (Adhikary et al., 2002). Hence, glaciers with a thick debris cover will have much lower ablation rates than ‘clean’ glaciers, as highlighted by Pelto (2000), who identified that annual ablation was significantly reduced by 25-30% on debris-covered ice, with summer ablation (once all snow cover has been lost) 30-40% less under debris-covered regions.

The impact of a debris layer on ablation rates means that debris-covered glaciers respond differently to a given set of climate conditions than clean glaciers (Jansson and Fredin, 2002). Debris-covered glaciers termini have often been noted as being unresponsive to climatic variations (Smiraglia et al., 2000), or that the fluctuations of the glacier termini are subdued (Kirkbride, in press). The glacier volume does, however, adjust in relation to climate, but unlike ‘clean’ glaciers these losses are redistributed throughout the glacier rather than being concentrated at the terminus, due to the protection of a thick debris cover (Thomson et al., 2000; Deline and Orombelli, 2005; Kirkbride, in press).

The factors outlined above illustrate that there is a requirement for improved monitoring of debris-covered glaciers to enable an increased understanding of their characteristics, dynamics and response to climate change. However, one limitation is that most glaciers occur in remote locations where access is difficult and the climate extreme (Rees, 2005), making field based
studies in many cases impossible. The remote sensing of debris-covered glaciers, therefore, provides a potential solution to this problem, making it possible to analyse glaciers in remote mountain locations and to monitor a large number of glaciers at the same time, at a high temporal frequency and on a global scale (Kaab, 2002; Bolch and Kamp, 2005; Kargel et al., 2005).

A number of studies have looked into glacier monitoring using remote sensing. However, the number of studies to debris-covered glaciers is significantly lower than those on ‘clean’ glaciers. Previous approaches on debris-covered glaciers have also been limited to those which aim to; i) identify their boundaries (e.g. Taschner and Ranzi, 2002; Ranzi, et al., 2004), ii) monitor debris cover changes over time (e.g. Stokes et al., 2007; Bolch et al., 2008; Shukla et al., 2009), iii) identify the rate of ablation under a debris layer using surface temperature measurements (e.g. Nakawo and Young, 1981; Nakawo et al., 1993; Nicholson and Benn, 2006), iv) surface elevation changes (e.g. Kaab, 2002; 2007) and velocity changes (e.g. Luckman et al., 2007)

Remote sensing has the potential to provide basic data required in the modelling of debris-covered glacier’s mass balance, hydrology and glacial dynamics. This potential can be utilised through the acquisition of data (from remotely sensed images) such as surface temperature, which enables the extraction of information for the monitoring of key topic areas including; debris cover extent, maps of debris thickness, and knowledge of the lithological type of a debris cover (which affects emissivity and thermal conductivity).

1.2. Aims and objectives

This project had one key aim:

To test the utility of visible/infrared satellite sensors with frequent repeat global coverage at a high temporal resolution, such as TERRA ASTER and Landsat (MSS, TM, ETM+), for studying debris-covered glaciers.
This aim was addressed through the development of techniques to fully exploit the capability of these satellite sensors to extract useful information. Consequently, four objectives were identified:

1) To develop a method to estimate debris thickness from ASTER thermal-band imagery using a physically-based energy balance model.

The identification of debris thickness is an essential process to determine the sub debris melt rates over glacier wide areas (Mihalcea et al., 2008). Previous attempts to achieve this have focused upon either field based measurements alone or empirical approaches (e.g. Mihalcea et al., 2008). Limitations to a field based method include the problem of spatial sampling of debris thickness, and the collection of enough points to interpolate a sufficiently accurate debris thickness map.

Remotely sensed estimates of debris covers have tended to use empirical approaches based on the relationship between debris thickness and surface temperature measured on the glacier on a single day (e.g. Mihalcea et al., 2008). However, the potential transferability in space and time of these empirical approaches is doubtful. The first objective, therefore, focuses on the development of a method with potentially increased transferability over previous field based and empirical approaches using an energy balance approach, which utilises limited and widely available meteorological input data variables and other parameters, which should be applicable to any glacier.

2) To use TERRA ASTER and Landsat data (MSS, TM, ETM+) to map changes in debris cover extent over time, using both manual and semi-automatic approaches.

An increase in debris cover has a direct impact upon the ablation of ice beneath. Therefore, the investigation and monitoring of debris cover extent is essential to determine any changes in melt
from debris-covered glaciers, which in turn impacts the runoff in melt water streams that provide a water resource for many in mountain regions (Berthier et al., 2004; Mihalcea et al., 2008). Remote sensing provides an under-used opportunity for mapping change in debris extent, with both a manual approach and semi-automatic approach (based on Paul et al., 2004) applied to assess their advantages and disadvantages.

3) To use ASTER DEMs and visible imagery to monitor both changes in surface elevation over time, and estimate surface velocities using feature tracking methods.

Changes in surface elevation provide a means of monitoring mass balance changes on glaciers (Etzelmuller, 2000; Racoviteanu et al., 2007), with different patterns of thickening and thinning expected on debris-covered glaciers compared with debris free, due to the influence of the debris layer (and its variation in thickness) upon surface ablation. The utilisation of remote sensing for this task enables the monitoring of an increased number of glaciers which previously have limited DEM coverage. Consequently, impacts upon a glacier’s surface elevation, surface velocity and mass balance due to climate changes can be identified and investigated at a frequent temporal resolution. Therefore, this objective used ASTER derived DEMs and visible imagery (orthorectified ASTER bands 1-3) to map elevation change and surface velocity.

4) To map the distribution of rock types in the supraglacial debris cover using spectral information derived from laboratory and satellite based remote sensing techniques.

The identification of rock types present on a debris-covered glacier is an area where remote sensing has significant potential due to the spectral differences of different rock types. The ability to identify rock types using remote sensing provides a solution to the limitation of field based methods (e.g. Deline, 2002) which are very time consuming. This objective applied ASTER data to identify different rock types present on the debris-covered surface based on their spectral signatures. In turn, the different emissivity values of these rock types were considered,
and the resulting impacts upon the estimation of surface temperature from satellite images highlighted.

1.3. Thesis synopsis

Chapter 2: An introduction to debris-covered glaciers and the use of remote sensing to monitor them along with an introduction to energy balance modelling on glaciers.

Chapter 3: Includes information on the study site of Miage Glacier including its location and key features. Justification for the selection of the Miage glacier as a study site and for the field data methods utilised is also provided.

Chapter 4: Provides information on the processes applied during this study and the methods employed to collect field data during a number of field campaigns, along with the satellite imagery utilised and any pre-processing (radiometric, geometric and atmospheric correction) applied identified.

Chapter 5: The first of four research chapters focused upon the development of an energy balance-based model for estimating debris thickness on debris-covered glacier surfaces using a surface temperature image from ASTER data. Stages in the model development are discussed along with any limitations which were identified and where possible overcome. The testing of empirical approaches, including that of Mihalcea et al., (2008), for debris thickness estimation was also investigated.

Chapter 6: This next research chapter investigated the potential of both manual and semi-automatic approaches for extracting debris extent using both Landsat (MSS, TM, and ETM+)
and ASTER images, and whether debris cover extent on the Miage Glacier has varied between 1975 and 2004.

**Chapter 7:** Two main foci were addressed in this chapter. Firstly, ASTER DEMs were utilised to identify surface elevation changes on the Miage Glacier between 2000 and 2006, although problems with the accuracy of these DEMs were encountered due to the steep and complex topography at the study site, which in turn limited analysis. Secondly, the surface velocity of the Miage Glacier was measured using feature tracking, however, DEM errors restricted result analysis.

**Chapter 8:** This final research chapter looked at the use of ASTER data for identifying the rock types present on a debris-covered glacier’s surface. Supervised classification was utilised for this, along with training data collected at Miage Glacier in 2007. The implications of different rock types and their emissivity values and thermal conductivities upon the calculation of surface temperature from thermal imagery (specifically ASTER AST08) were also investigated.

**Chapter 9:** Provides a discussion into what was known prior to the completion of this project, and what is now know after its completion, highlighting how gaps in the research on debris-covered glaciers have (or have not) been filled. Areas highlighted as requiring future work during the completion of this study are also discussed, along with the identification of the applicability of the findings of this study from both a glaciological and remote sensing perspective.

**Chapter 10:** This conclusion chapter highlights the importance of the research topic and provides a summary of the main findings of this research project.
CHAPTER 2: APPLICATION OF REMOTE SENSING TO THE STUDY AND MONITORING OF DEBRIS-COVERED GLACIERS: A LITERATURE REVIEW

2.1. Introduction

Remote sensing has the potential to provide a variety of useful environmental information relating to debris-covered glaciers, although its use so far remains limited. This chapter summarises current understanding of debris-covered glaciers and reviews both the principles and application of remote sensing methods, and their potential to provide information suitable for incorporation into a variety of monitoring techniques. These include debris thickness estimation, surface elevation changes, debris extent variations and identification of debris cover lithologies.

2.2. Debris-covered glaciers

The addition of debris to valley glaciers is dominated by both rockfalls and avalanches from mountain faces, especially in temperate alpine regions (Kirkbride, 1995). It may also be deposited on a glacier surface by windfall (particularly volcanic ash, dust), and human action (Bennett and Glasser, 1996). The transport of this material on the glacier’s surface is termed high-level transport. The amount and variety of sediment material which enters this high-level (supraglacial) transport is dependent upon both the nature and extent of the extra-glacial terrain, and also the effectiveness of weathering and erosion upon this (Kirkbride, 1995). Debris can also originate from the bed as a result of subglacial erosion and englacial thrusting to the surface (Owen et al., 2003). Glaciers can be partly or continuously covered by debris, and those where debris covers the full width of part of the ablation zone may be defined as debris-covered (Singh et al., 2000; Nakawo et al., 1993; Mattson et al., 1993). However, some authors specify that to
be a debris-covered glacier, 50% of the ablation zone needs to be covered (Kirkbride, in press).

It is also important to note that this supraglacial debris cover can exhibit significant spatial variability in thickness, grain size, and mineralogy, which reflect the distribution of sources, transport paths, and subsequent reworking of debris on top of and within the glacier (Benn and Evans, 1998).

Material in high-level debris transport is exposed to highly active weathering processes including freeze thaw and chemical weathering, which is a consequence of an abundant moisture supply (Bennett and Glasser, 1996). Also, if the supraglacial debris layer is stable, biological weathering may also take place as plants and other vegetation colonise the debris surface. This results in debris material being broken down from larger clasts to fines, and due to the nature of the weathering processes these particles remain angular and do not become rounded. Large clasts may also be broken down through the crushing of large boulders by inter-particle stress, caused by the weight of large boulders on top of smaller boulders. This process has been observed on the Miage Glacier in Italy, where the grain crushing results through inter-boulder contact in both the englacial and supraglacial environment (Owen, et al., 2003). Therefore, its debris characteristics (coarse and angular) are similar to those of talus or scree slopes which reflect the common origin of both of these as rockfall debris (Bennett and Glasser, 1996).

If rockfall material falls onto the glacier surface above the equilibrium line, it becomes buried beneath successive layers of snow accumulation and becomes concentrated into layers and incorporated into the englacial transport system (Owen, et al., 2003). It is then frequently either re-exposed in the ablation zone, when surface ice melts (Figure 2.1), or forced to the surface by englacial thrusts of ice (Benn and Evans, 1998). If it falls onto the glacier surface below the equilibrium line it will stay exposed on the glacier surface until it reaches the snout, where
debris transfer occurs due to compressive flow (Bennett and Glasser, 1996), whilst undergoing mass movement processes and re-distribution by melt water.

Figure 2.1: Exposure of a medial moraine in the ablation zone, AD = ice-debris interface, AE = moraine crest, B = ice core, AC = level of nearby glacier not affected by differential ablation (Kirkbride, 1995).

On the other hand, if debris originates from the bed, or comes into contact with it during transport (termed low level debris transport) (Bennett and Glasser, 1996) the particles’ size and shape are altered significantly. As a result, the debris material consists of fine grained materials which are rounded and not angular (Bennett and Glasser, 1996; Owen et al., 2003). Therefore, the form and particle size distribution of debris material on a glacier surface is dependent upon both the transport pathway which the material has travelled (Owen, et al., 2003) and, more importantly, the distance from its source.
2.2.1. Development of debris cover and lateral debris movement

On a glacier’s surface, the rate of ablation is strongly influenced by the distribution and thickness of any debris material, with troughs developing where debris is thin and ablation is maximised. Compared to areas where thicker debris is present, where ablation levels are reduced protecting the ice beneath, and results in higher ridge areas on a glaciers surface (Kirkbride, 1995).

![Figure 2.2 Cycle of ridge and trough development.](image)

The development of ridges and troughs is a continuous process because once troughs have formed, debris will fall from the steep sided ridges into the troughs under gravity, building up a thicker debris layer which protects the ice from ablation. Therefore, over time the debris cover on the ridges becomes thinner and ablation greater, and as a consequence the height of these
ridges above the trough areas is gradually lowered, until they become troughs beneath debris-covered ridges (previously the troughs) where the cycle repeats again (Figure 2.2).

Such development was observed by Pelfini et al., (2007), who noted that the debris-covered lobes of the Miage Glacier appear as undulating in a variable way, resulting from the presence of niches, depression zones and channels, a morphology which results as a consequence of differential ablation. The presence of medial moraines has the same impact as thicker debris-covered areas and lowers ablation levels. These medial moraines become increasingly exposed down glacier as the ice surrounding them melts due to greater melt rates in the ablation area (Figure 2.1). This results in the development of large ridges of debris on a glacier surface, characterised by increasing surface relief (Kirkbride and Warren, 1999). Also, the lateral movement of debris from the moraine increases the area affected by differential ablation (Kirkbride, 1995).

### 2.3. Impact of debris cover on glacial surface processes

The presence of debris on a glacier surface results in the modification of a number of surface processes. Firstly, the thermal regime of the surface is affected and, in turn, this influences the process of ablation beneath this debris layer.

#### 2.3.1. Thermal Properties of a debris layer

A debris layer overlying glacier ice hinders the heat flow from the top of the debris surface to the ice underneath and acts as an insulator (Nakawo and Rana, 1999), with energy released to the atmosphere during the day (when the debris is hottest) instead of being conducted downward and melting the ice. The ability of a debris layer to store heat and act as an insulator is determined by its thermal properties. Consequently, the thermal properties of the debris layer will control its surface temperature. However, it must also be noted that the surface temperature...
is not just a function of the thermal properties of the debris, but also a function of altitude, surface orientation, surface slope, and the exchange of heat between the surface and the surrounding atmosphere (Nakawo et al., 1993). The important thermal properties of a debris layer are thermal conductivity, thermal resistance, thermal diffusivity, and the specific heat capacity.

### 2.3.1.1. Thermal conductivity ($K$)

Thermal conductivity ($K$) is a key thermophysical property of rocks (Abdulagatov et al., 2006) and is a measure of the rate at which heat flows through a certain thickness of material (i.e. its ability to transmit heat) (Jumikis, 1977; Kahle, 1980; Oke, 1987). An increase in conductivity will result in the increased penetration of heat into the debris layer, which in turn results in a lower rise in temperature at the surface (Kahle, 1980). Therefore, it is the main physical characteristic of a debris layer that controls heat conduction to the debris-ice interface and can control the rate of heat transfer into the ice below, with different materials having very different conductivity values (Table 2.1). Where: $COND$ (conductive heat flux $W \, m^{-2}$), $d$ (debris thickness, $m$), $T_s$ (surface temperature $^\circ C$), and $T_{di}$ (Temperature at debris-ice interface $^\circ C$):

\[
K = \frac{COND - d}{T_s - T_{di}} \tag{2.1}
\]

<table>
<thead>
<tr>
<th>Material</th>
<th>Thermal conductivity (Wm$^{-1}$ K$^{-1}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Basalt$^1$</td>
<td>2.09</td>
</tr>
<tr>
<td>Granite$^1$</td>
<td>3.32</td>
</tr>
<tr>
<td>Slate$^1$</td>
<td>2.09</td>
</tr>
<tr>
<td>Shale$^1$</td>
<td>1.76</td>
</tr>
<tr>
<td>Limestone$^1$</td>
<td>2.00</td>
</tr>
<tr>
<td>Sandstone Quartz$^1$</td>
<td>5.02</td>
</tr>
<tr>
<td>Snow$^2$</td>
<td>0.08</td>
</tr>
<tr>
<td>Ice$^2$</td>
<td>2.24</td>
</tr>
<tr>
<td>Water$^2$</td>
<td>0.57</td>
</tr>
<tr>
<td>Air$^2$</td>
<td>0.03</td>
</tr>
</tbody>
</table>

Table 2.1: Thermal conductivities of a variety of materials$^1$Gupta (2003),$^2$Oke (1987)
Where the thermal conductivity of debris is constant in a profile, a linear thermal gradient exists (Jumikis, 1977). However, discontinuities within a debris layer mean that thermal conductivity can vary within a profile and results in a non-linear thermal gradient. In particular, water can affect the thermal gradient within a profile as the introduction of water replaces the less conductive air from interstitial spaces within the debris layer (Jumikis, 1977; Oke, 1987). The presence of water is increasingly common towards the bottom of the debris layer at the debris-ice interface due to melting ice. Therefore, the thermal conductivity towards the bottom of a debris profile (at the ice interface) will vary to that at the surface due to the higher conductivity of water (Brock et al., 2007).

### 2.3.1.2. Thermal resistance (Tr)

The thermal resistance (Tr) of a debris layer is equal to the debris thickness divided by its thermal conductivity (Jumikis, 1977; Nakawo et al., 1993) (Equation 2.2). Thermal resistance is, therefore, the resistance of a material (of a certain thickness) to the penetration of heat (which is dependent upon its thermal conductivity) (Jumikis, 1977):

\[
Tr = \frac{d}{K}
\]  

(2.2)

Thin debris layers have a lower thermal resistance due to there being less material for heat to be dissipated into. Thicker debris layers have a larger thermal resistance as a greater amount of heat is dissipated and stored within the debris (generating a high debris surface temperature) and not transported through to the debris-ice interface (Nakawo and Young, 1981; Mattson, 2000).

### 2.3.1.3. Thermal diffusivity (Tdf)

Thermal diffusivity (Tdf) (also known as the temperature conductivity or thermometric conductivity) is related to the variable flow of heat through a material (Jumikis, 1977). It is, therefore, a measure of the rate at which a change in temperature at the surface is spread
throughout the material beneath it, with values varying for different materials (Table 2.2).

Materials with high diffusivities allow the rapid penetration of surface temperature changes to the layers below, meaning temperatures can be altered over a thicker layer. This provides an index of the ease with which a material undergoes a temperature change. Therefore, the thermal diffusivity of a debris is directly proportional to its ability to transmit heat, but is inversely proportional to the amount of heat that is required to generate a temperature change (specific heat capacity) (Oke, 1987). Where; $c$ (Specific heat capacity $804 \text{ J K}^{-1} \text{ Kg}^{-1}$):

$$Tdf = K/c \quad (2.3)$$

<table>
<thead>
<tr>
<th>Material</th>
<th>Thermal diffusivity $(\text{m}^2 \text{s}^{-1} \times 10^{-6})$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Basalt$^1$</td>
<td>0.9</td>
</tr>
<tr>
<td>Granite$^1$</td>
<td>1.6</td>
</tr>
<tr>
<td>Slate$^1$</td>
<td>1.1</td>
</tr>
<tr>
<td>Shale$^1$</td>
<td>0.8</td>
</tr>
<tr>
<td>Limestone$^1$</td>
<td>1.1</td>
</tr>
<tr>
<td>Sandstone Quartz$^1$</td>
<td>1.3</td>
</tr>
<tr>
<td>Snow$^2$</td>
<td>0.1</td>
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<td>1.2</td>
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<td>0.1</td>
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<tr>
<td>Air$^2$</td>
<td>20.5</td>
</tr>
</tbody>
</table>

Temperature regimes in thicker debris layers tend to be less extreme because the surface heating during the day warms the thick layer, and any night time surface cooling is offset by the heat being drawn up from lower layers (Oke, 1987). More extreme diurnal temperature fluctuations result from thinner debris layers.

### 2.3.1.4. Specific heat capacity ($c$)

The specific heat capacity ($c$) determines the amount of heat (Joules) needed to raise the temperature of a unit volume ($\text{m}^3$) of a material by 1 degree (Kelvin) (Oke, 1987). Therefore, a substance with a high specific heat capacity will absorb more heat before its temperature is
raised, preventing the heat from being transported into the ice below. The specific heat capacity of a substance is influenced by the presence of water which requires a large input of heat to cause certain change in temperature, compared to air which requires significantly less (Table 2.3) (Oke, 1987).

Table 2.3: Example Specific heat capacities \(^1\)Gupta (2003), \(^2\)Oke (1987)

<table>
<thead>
<tr>
<th>Material</th>
<th>Specific heat capacity (J Kg(^{-1}) K(^{-1}))</th>
</tr>
</thead>
<tbody>
<tr>
<td>Basalt(^1)</td>
<td>837</td>
</tr>
<tr>
<td>Granite(^1)</td>
<td>669</td>
</tr>
<tr>
<td>Slate(^1)</td>
<td>712</td>
</tr>
<tr>
<td>Shale(^1)</td>
<td>712</td>
</tr>
<tr>
<td>Limestone (^1)</td>
<td>712</td>
</tr>
<tr>
<td>Sandstone Quartz(^1)</td>
<td>712</td>
</tr>
<tr>
<td>Snow(^2)</td>
<td>2090</td>
</tr>
<tr>
<td>Ice(^2)</td>
<td>2100</td>
</tr>
<tr>
<td>Water(^2)</td>
<td>4180</td>
</tr>
<tr>
<td>Air(^2)</td>
<td>1010</td>
</tr>
</tbody>
</table>

2.3.1.5. **Conductive heat flux (COND)**

The conductive heat flux represents the total amount of heat transferred through a material, with the temperature at a location, thermal conductivity, and debris thickness required to complete the calculation (Equation 2.4). However, this assumes that thermal conductivity is constant with depth, whereas it probably increases with depth as the air spaces (lower conductivity) become filled with water which has a higher conductivity (Brock *et al.*, 2007). The conductive heat flux is a key component in surface energy balance modelling, due to it being a measure of the amount of heat which can be conducted through a debris layer and used to melt the ice below.

\[
COND = K \frac{T_s}{d} \tag{2.4}
\]

2.3.2. **Factors which affect the thermal properties of a debris layer**

A number of factors influence the thermal properties of a debris layer and, in turn, have an impact upon the resulting ablation rates under a debris layer on a debris-covered glacier. These include sediment composition, lithology, and the moisture content of the debris layer.
2.3.2.1. Sediment texture (grain size)

Different thermal regimes exist between coarse blocky materials and finer mineral soils, resulting in variable ablation rates (Harris and Pedersen, 1998). Fine-grained debris covers have been identified as providing significantly better insulation from ablation conditions (Pelto, 2000). This may result due to two reasons. Firstly, porous fine-grained debris particles may have air trapped between them; and because air has a much lower conductivity than water it results in the reduction of the thermal conductivity of the debris layer and, therefore, reduced ablation rates (Jumikis, 1977). Secondly, fine-grained debris particles have lower thermal diffusivity rates due to the lack of interconnecting voids, which in turn, results in a lower thermal conductivity (Harris and Pederson, 1998). Warmer temperatures are experienced in the surface layers of finer grained debris, due to its insulating effect which, as a result, increases the debris temperature as it absorbs and retains more heat than coarse blocky material (Gorbunov et al., 2004). However, colder temperatures will be experienced at lower levels in the debris profile by the finer grained material, a consequence of the finer sediments closer proximity to the ice below.

Coarse blocky material absorbs and retains less heat due to the rapid air movement through the large interconnecting voids, and because the large blocky material contains fewer pore spaces for warmer air to be retained (Harris and Pedersen, 1998). As a result, the thermal response to a change in air temperature can be more immediate, with hotter or cooler air temperatures transported through the debris layer much faster than in finer grained debris covers, leading to a more variable thermal regime.

2.3.2.2. Sediment lithology (composition)

The lithology of a debris layer also plays a key role in the amount of ablation that occurs. Albedo of the debris layer can have an important influence on the conductivity of a debris layer
and the resulting ablation. Debris layers composed of dark schists, for example, increase \textit{COND} by increasing the temperature of a surface, leading to much higher levels of ablation than a debris layer of lighter coloured granite (Conway and Rasmussen, 2000). This results because dark-coloured materials absorb solar radiation much better than light-coloured materials. As a result the \textit{COND} of darker-materials is higher and results in greater ablation (Table 2.4a) (Jumikis, 1977). As well as the sediment lithology having a key impact upon albedo, Brock \textit{et al.}, (2000) highlight that the percentage of debris cover can have a significant impact (Table 2.4b), with a greater percentage of debris cover typically having a lower albedo. This is a result of less ice being visible which has a much higher albedo compared to the debris material.

<table>
<thead>
<tr>
<th>Table 2.4: a) published albedo values for different surfaces, \textsuperscript{1} Knap, \textit{et al.}, (1999), \textsuperscript{2}Jumikis (1977), \textsuperscript{3} Oke (1987), \textsuperscript{4} Gupta (2003,) b) variation of ice albedo with differences in debris cover (Brock \textit{et al.}, 2000)</th>
</tr>
</thead>
<tbody>
<tr>
<td>a)</td>
</tr>
<tr>
<td>Material</td>
</tr>
<tr>
<td>Snow – fresh\textsuperscript{2}</td>
</tr>
<tr>
<td>Snow – old\textsuperscript{2}</td>
</tr>
<tr>
<td>Slate \textsuperscript{3}</td>
</tr>
<tr>
<td>Limestone\textsuperscript{4}</td>
</tr>
<tr>
<td>Granite\textsuperscript{4}</td>
</tr>
<tr>
<td>b)</td>
</tr>
<tr>
<td>Debris cover (%)</td>
</tr>
<tr>
<td>0</td>
</tr>
<tr>
<td>20</td>
</tr>
<tr>
<td>60</td>
</tr>
<tr>
<td>80</td>
</tr>
<tr>
<td>100</td>
</tr>
</tbody>
</table>

2.3.2.3. \textit{Debris moisture content}

The meteorological conditions at a site can also influence the thermal properties of a debris layer. For example, a three fold increase in the percentage of absorbed surface energy reaching the debris/ice interface was observed on the Rakhiot glacier, Pakistan, when the debris layer became moistened by rainfall (Mattson, 2000). Therefore, a greater proportion of energy can be
transferred into the debris layer below when the debris is wet due to the high conductivity of water, and resulting increase in thermal conductivity (Jumikis, 1977).

### 2.3.3. Ablation rates on debris-covered glaciers

Ablation can be defined as the rate of decrease of the water-equivalent thickness of ice/snow due to mass loss (Patterson, 2001). The assessment of the ablation rate under a debris layer can provide information for many purposes, including the investigation of glacier mass balance and the impact of climate change upon glacial melt rates, glacier dynamics and glacial history investigations (Nakawo and Young, 1981). The presence of debris material has a significant impact upon ablation rates (Table 2.5), with the amount of ablation dependent upon the debris thickness (Figure 2.3, Table 2.6).

**Table 2.5: Comparison of mean annual ablation rates (between 1984-1998) on clean and debris-covered ice on the Columbia Glacier and Lyman Glacier, North Cascades, Washington (Pelto, 2000)**

<table>
<thead>
<tr>
<th>Glacier</th>
<th>Annual ablation: clean glacial ice</th>
<th>Annual ablation: debris covered ice</th>
</tr>
</thead>
<tbody>
<tr>
<td>Columbia Glacier</td>
<td>3.3 m water equivalent</td>
<td>2.3 m water equivalent</td>
</tr>
<tr>
<td>Lyman Glacier</td>
<td>3.4 m water equivalent</td>
<td>2.6 m water equivalent</td>
</tr>
</tbody>
</table>

**Table 2.6: Critical and effective thicknesses, \(^1\text{Mattson (2000), }^2\text{Popovnin and Rozova (2002), }^3\text{Mattson et al., (1993), }^4\text{Kayastha et al., (2000)}\)**

<table>
<thead>
<tr>
<th>Location</th>
<th>Critical thickness (mm)</th>
<th>Effective thickness (mm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Dome Glacier, Canadian Rockies(^1)</td>
<td>20</td>
<td>-</td>
</tr>
<tr>
<td>Djankuat Glacier, Caucasus(^2)</td>
<td>70-80</td>
<td>-</td>
</tr>
<tr>
<td>Lirung Glacier, Nepal(^3)</td>
<td>1.3</td>
<td>0.25</td>
</tr>
<tr>
<td>Khumbu Glacier, Nepal(^4)</td>
<td>-</td>
<td>3</td>
</tr>
</tbody>
</table>

Two key terms are used when analysing the impact of debris upon ablation and these are the critical thickness and the effective thickness. The critical thickness is the thickness at which sub-debris ablation rate equals that for clean ice/snow (Kayastha et al., 2000). Therefore, the key role of debris thickness in controlling ablation is highlighted, and if the debris cover on a glacier...
is thicker than its critical thickness, the ice beneath will be insulated and ablation significantly reduced. The effective thickness is the debris thickness beneath which ablation is maximised and will be greater than that experienced on bare ice, highlighting the negative impact of thin debris layers upon a glacier surface (Kirkbride and Dugmore, 2003). The variability of these values from glacier to glacier is highlighted by Figure 2.3 and Table 2.6.

![Figure 2.3: Relationship between debris cover thickness and ablation (Mattson et al., 1993).](image)

2.4. Thermal remote sensing

Irradiance is the radiant flux of energy incident upon a surface, whereas radiance is the radiant flux of energy leaving a surface area in a given direction, and thermal sensors record radiance (Wm\(^{-2}\) sr\(^{-1}\)) (Jensen, 2000). Thermal remote sensing has considerable potential for the application of monitoring debris-covered glaciers by providing surface temperature data. This enables the identification of debris-covered glacial boundaries and debris cover extent, due to the lower surface temperature of debris-covered ice (except when debris is > 0.5 m). Both glacial boundaries and debris extent are difficult to determine with visible imagery alone due to
the presence of moraines, which are visually very similar to debris-covered glacial ice (Paul et al. 2004). Surface temperature data can also be used to estimate thermal resistance (and thickness if thermal conductivity is known).

### 2.4.1. Thermal infrared energy

Infrared energy adjoins the red end of the visible part of the electromagnetic spectrum and comprises three different categories, thermal infrared (3-14 µm), mid infrared (1.3-3 µm), and near infrared (0.7-1.3 µm). Thermal infrared energy differs slightly to the near and mid infrared regions as only thermal energy is directly related to the sensation of heat and the radiation of energy from an object. It is termed ‘thermal infrared’ because this type of radiation is correlated with the temperature of the emitting surface or object (Lillesand et al., 2004). Therefore, a dividing line exists between the reflected and emitted infrared radiation, which occurs at ~3 µm. Above 3 µm emitted energy dominates (heat) and below 3 µm reflected energy dominates. All surfaces emit thermal radiation when their temperatures are above absolute zero (Jensen, 2000), and the amount of energy radiated by an object is a function of its surface temperature (Kahle, 1980).

![Absorption, emission, transmission, and reflection of thermal infrared energy from an object.](image)

**Figure 2.4:** Absorption, emission, transmission, and reflection of thermal infrared energy from an object.
The thermal infrared energy recorded by a remote sensor consists of energy emitted from an object. However, the energy which is radiated from the object is added to by radiation reflected from the surface. Some of this energy is absorbed by the object (absorptivity), and some is transmitted by the object, and the rest is reflected back along with the radiation emitted from the object itself (Figure 2.4). The temperature of an object (and the resulting thermal infrared radiation received) and any extremes/variations of this are determined by three main factors: thermal conductivity, thermal capacity and thermal inertia.

### 2.4.2. Interaction of thermal infrared energy with the atmosphere

Once electromagnetic radiation is generated, it travels through the Earth’s atmosphere, which can affect a number of its properties, such as its wavelength, intensity and spectral distribution. Its path through the atmosphere can also be altered due to refraction (Jensen, 2000). The atmosphere, therefore, has a significant effect upon the intensity and spectral composition of the energy which is received and recorded by a thermal system.

The atmosphere can either increase or decrease the amount of radiation coming from the ground, and the extent of the atmospheric impact upon the received signal depends upon the degree of atmospheric absorption, scattering and emission at the time and location of sensing (Lillesand et al., 2004). Atmospheric scattering (Rayleigh, Mie, and non-selective scattering) and absorption (most common components being: water vapour, carbon dioxide, and ozone) make objects on the ground appear much colder than they actually are, and emission from the atmosphere makes the objects appear much warmer.

The overall effect of the atmosphere varies depending on a number of factors including the path length (distance between the ground and the sensor), magnitude of the signal (weaker signal will be influenced more than a stronger signal), the atmospheric conditions present at the time of sensing.
monitoring (the presence of cloud increases scattering and absorption, as does rain), and the wavelengths that are being sensed, because at some wavelengths the atmosphere completely blocks all radiation (Lillesand et al., 2004). As a result, sensors are designed to record information in atmospheric windows such as those at 3-5 μm and 8-14 μm.

2.5. Previous applications of satellite imagery: ‘clean’ glacial monitoring

Remotely sensed data have been used for a variety of ‘clean’ glacial applications. These have been focused mainly on the determination of surface albedo, the monitoring of glacier motion, and the monitoring of changes in the mass balance of glaciers, ice sheets and ice shelves. All of these approaches have been utilized by the Global Land Ice Measurements from Space (GLIMS) programme and a brief introduction to this is included.

2.5.1. GLIMS

The Global Land and Ice Measurements from Space project (GLIMS) was initiated as a glacial inventory, which is updated on a regular basis and provides a useful resource both to scientists and policy makers in solving problems, which may result from any changes or fluctuations of the Earth’s climate (Ranzi et al., 2004). GLIMS is an international consortium which was established to acquire satellite images of the world’s glaciers, analyse them and identify any changes, and highlight whether any changes that have occurred will affect people and the environment (Kargel et al., 2005). The project consists of 3 main goals (Raup et al., 2000):

1. To acquire ASTER image data covering all of Earth’s permanent land ice once per year in the local melt season.
2. To derive glaciological information from the imagery using automated computer techniques.
3. Archive the results in a widely accessible geographic computer system using a GIS.
The generation of a global glacial inventory enables the production of increasingly detailed Global Climate Models (GCM) which aim to predict/model future trends in climate. As more known factors are incorporated into them, such as glacier dynamics and changes over time, the more accurate the output predictions will be (Kargel et al., 2005). In total, the GLIMS project will use satellite data (mainly ASTER and Landsat) to map and catalogue approximately 80,000 glaciers (Schmugge et al., 2002).

### 2.5.2. Surface albedo

The albedo of a surface is a measure of the ratio of reflected to incident radiation (in the 0.3-3 \( \mu \text{m} \) range). Both snow and ice have a much higher albedo compared to other surfaces, with temporal and spatial variations of albedo being the dominant control on melt rates on glaciers (Greuell and de Ruyter de Wildt, 1999; Knap et al., 1999). Lower albedo values result in increased melt rates as more shortwave radiation is absorbed and used for melting (Gruell and de Ruyter de Wildt, 1999). Remote sensing studies concerned with surface albedo measurements on glaciers are focused upon the calculation of the radiation balance from Landsat TM imagery, from which surface albedo is calculated. Further details of the methods (components calculated and Equations applied) and variation in these can be found in Hall et al., (1990), Knap et al., (1999), and Greuell and de Ruyter de Wildt (1999).

The advantage of using remote sensing to determine albedo occurs because ground based measurements are not always representative of a large area due to the varying spatial distribution of albedo on a variety of scales (Greuell and de Ruyter de Wildt, 1999). Satellite measurements of albedo are compared to ground albedo measurements for validation (Hall et al., 1990; Knap et al., 1999), with comparisons highlighting that satellites tend to over estimate the albedo of both snow and ice and this results for a number of reasons. Firstly, the spatial resolution of pixels which provides an average of a 900 m\(^2\) area (for Landsat), meaning variation in albedo over this
area is not considered (Knap et al., 1999). Secondly, problems of poor temporal resolution of the satellite data relative to the rate of albedo change, because, snow albedo may decrease rapidly in a few hours following a snowfall (Knap et al., 1999). Finally, overestimation results due to the satellite viewing at nadir, because the full spectral reflectance is greater off nadir due to forward scattering (Hall et al., 1990). To correct this, bi-directional reflectance models can be utilized (Greuell and de Ruyter de Wildt, 1999).

2.5.3. Glacier motion

To determine glacier motion using remote sensing, Synthetic Aperture Radar (SAR) interferometry is often used, where two side looking radar images of the same scene are taken from the same point in satellite orbit but at different times are combined together to produce an interference map (interferogram). If the surface of the glacier has changed between view’s it will show on the map as interference (essentially a phase shift or measure of displacement of a ground surface point in relation between two images) (Goldstein et al. 1993; Luckman et al., 2007). From the movement observed over a given period velocity can be determined. Results obtained using this method on glaciers in the French Alps have shown to be as accurate as terrestrially obtained velocity measurements (Bousquet et al., 2004).

An alternative approach to SAR interferometry is a feature tracking approach where the displacement of features between two (or more) satellite images is measured, and from which surface velocities can be calculated. To complete this processes surface features have to be detectable in at least two of the repeat data sets, to enable their tracking. Also the multi-temporal datasets must be accurately co-registered to avoid mis-registration errors which can be misinterpreted as movement. Finally, the spatial resolution of the datasets utilised must be finer than the surface displacements that occur on the glacier (Kaab, 2005). Kaab (2005) successfully
applied a feature tracking approach combining ASTER and SRTM data in the Bhutan Himalaya, and highlighted the potential of spaceborne optical data for obtaining velocity measurements.

2.5.4. Changes in glacial extent and elevation

Remote sensing has been used in studies looking into the changes in the areal extent and elevation of glaciers (Paul 2002; Thomas et al., 2004; Berthier et al., 2004; Paul et al., 2007), ice sheets (Thomas et al., 2009), and ice shelves (Frezotti, 1993). One example of a study on extent variations of Alpine glaciers is provided by Paul (2002), who analysed changes in areal extent of 235 glaciers between 1969-1992 in Tyrol, Austria. Using a band ratio on Landsat TM data (channels 4 and 5), and glacier mask from which glacier areas where calculated for each year (by counting the number of pixels and multiplying it by the scale of the Landsat pixels: 900 m²).

Paul (2002) found glaciers with an area less than 1 km² retreated significantly between 1969-1992 by ~35% and a total loss in the study area during this period of ~43 km². Paul et al., (2007) also monitored glaciers in Switzerland and identified downwasting using Landsat TM and ASTER data through the identification of features which indicate downwasting is occurring, including; growing rock outcrops, separation of the glacier from its tributaries, and the formation of pro-glacial lakes.

Satellite altimeters can also be utilised to determine surface elevation changes, and measure the elevation of a surface by measuring the time taken for a pulse of electromagnetic radiation to be sent from the satellite to the ground and back (Gibson et al., 2000). From this elevation changes can be calculated using multitemporal images by subtracting one image from another. Shepherd et al., (2001), for example, utilised European Remote Sensing (ERS) altimeter data from 1992
and 1997 during a study on Pine Island Glacier, West Antarctica, and found thinning rates of -1.6 m y\(^{-1}\).

LiDAR data has also been widely used to generate DEMs for glacial surface change investigations, where the time taken for a pulse of light from a laser transmitter to a target and back again is measured, and from this elevation can be determined (Jensen, 2000). To identify surface elevation changes, image subtraction is performed on LiDAR images from two different images. One example is provided by Kohler et al., (2007) who utilised LiDAR data during a study of thinning rates on western Svalbard glaciers. The elevational accuracy of LiDAR during this study was also tested and compared to independently collected GPS data, and was identified as being +/- 0.25 m.

**2.6. Previous applications of satellite imagery: debris-covered glacier monitoring**

Following an investigation into research undertaken on glaciers, it became clear that a significantly larger amount of studies had focussed upon ‘clean’ glaciers. Therefore, a key area requiring focus became apparent, and provided a basis for this study into increasing the possibilities of monitoring debris-covered glaciers. Both Landsat TM and ASTER satellite imagery have previously been used to monitor debris-covered glaciers. This has included the identification of debris-covered glaciers through boundary mapping, the generation of digital elevation models (DEM), surface velocity measurements, and estimation of debris thickness on the glaciers surface.

**2.6.1. Glacier boundary and debris mapping**

Initially, the mapping of debris-covered glacier boundaries was limited to manual delineation, which involves the process of manually digitising around a debris layer to identify its extent
(e.g. Stokes, et al., 2007). The dominance of manual delineation resulted due to the presence of a supra-glacial debris layer, which exhibits the same spectral properties of glacial moraines and, therefore, hinders the automatic or semi-automatic discrimination of glacial boundaries using visible imagery alone (such as band ratios e.g. Equation 2.5 and 2.6) (Paul et al., 2004; Ranzi et al., 2004). However, manual digitizing (although accurate) is both time-consuming and labour intensive when studying a large number of glaciers (Paul et al., 2004).

\[ \text{Landsat} = \frac{TM4}{TM5} \quad (2.5) \]

\[ \text{ASTER} = \frac{VNIR3}{SWIR4} \quad (2.6) \]

Taschner and Ranzi (2002), however, observed that debris overlying ice has a much lower temperature than extra glacial debris (temperature approx. 4.5 °C warmer) (Figure 2.5). Therefore, automatic methods which used band ratios on visible channels could be modified to include ASTER imagery, which provides further information through its thermal channels (Taschner and Ranzi, 2002; Ranzi et al., 2004; Paul et al., 2004). This (along with a NDVI image) allows the boundaries of a debris-covered glacier to be identified from its surroundings and, as a result, changes in both size and extent can be monitored over time. However, as this only detects the boundaries of a debris-covered glacier, other methods need to be applied to detect the extent of debris cover on a glacier’s surface and its variation over time.

Another issue which needs to be highlighted is that often the temperatures of both bare rock and debris lying on ice are identical at the time of ASTER image acquisition (10:40, Figure 2.5), which will increase the difficulties in making a discrimination based on surface temperature differences alone. Also, for debris overlying ice to have a much lower surface temperature than
bare rock, it needs to be thin enough to bare the thermal signature of the underlying ice, as well as being at a time of day when there is a temperature difference.

**Figure 2.5:** Diurnal surface temperature variations of both bare rock and debris lying on ice (Taschner and Ranzi, 2002).

An automated method to identify the extent of debris cover on a glacier surface was developed by Paul *et al.*, (2004) based on a band ratio approach. Using Landsat TM data, a ratio image was combined with a DEM, where all slopes $>24^\circ$ were excluded, because a 0-24 $^\circ$ range is not exceeded by many debris-covered glacier tongues (Paul *et al.*, 2004), and an intensity, hue, saturation (IHS) image of three of the Landsat channels (TM3, TM4, and TM5). By applying thresholds to each of the separate images (which identified the debris cover area most accurately from their surroundings), the images are combined together to produce a final debris extent map. This was developed and applied successfully to the Oberaletschgletscher in the Swiss Alps, with results comparable to a previously generated vector data set debris extent map generated by the Swiss Federal Office of Topography (Paul *et al.*, 2004). Any difference which occurred between the observed and estimated debris extent were attributed to the slope angle of $>24^\circ$. 
used, and changes in glacial extent since the debris-extent map was completed (1993). Where differences did occur between estimated and observed debris extent they could be corrected manually.

### 2.6.2. Digital elevation models (DEMs), generation, and surface change estimation

DEMs can be used to measure changes in elevation of the glacier surface (by subtracting one DEM from another), which can be compared over time to estimate changes in glacier volume, and consequently changes in surface melt rates, velocities, and ice inputs. These surface elevation changes can also be used to assess glacier mass balance changes over time (Racoviteanu et al., 2007). Previous studies on debris-covered glaciers include Diolaiuti et al., (2009) on Miage Glacier which utilised orthophoto DEM data to obtain surface elevation changes between 1975-2003.

ASTER DEMs have been previously utilised for this application, however, some limitations have resulted when DEMs have been generated from ASTER imagery, mainly the over-estimation of terrain heights, especially in the case of steep terrain (Kaab, 2002; Bolch and Kamp, 2005; Racoviteanu et al., 2007). For example, Kaab (2002) noted errors of +/- 60 m for rough high mountain topography, and +/- 18 m for more moderate mountainous terrain (Kaab, 2002) in Ruben, the Swiss Alps. Kaab (2002) also highlighted that DEM errors were greatest at sharp peaks with northern flanks, as they are hidden from the back looking stereo channel (Kaab, 2002). However, despite these limitations, the ASTER DEM product is a valuable tool for monitoring debris-covered glaciers in high mountain regions which have poor or limited elevation data (Paul et al., 2004).

The advantage of ASTER data for this application is its large global data set which has been acquired since 2000 (Kaab, 2007). Therefore, the impact of cloud (which obscured the ground
beneath, meaning accurate elevation values cannot be determined) is not as profound, as another image can be used from a later date where cloud cover is reduced, whereas, the coverage and reoccurrence of other DEM missions has been limited.

### 2.6.3. Surface velocity measurements

The monitoring of frontal recession is a poor indicator of negative mass balance on debris-covered glaciers, because, downwasting is the most common source of mass loss. An alternative to surface elevation change analysis is an investigation into glacier flow rates, because a reduction in velocities may identify stagnation of the glacier (Luckman et al., 2007). The monitoring of surface velocities of glaciers, therefore, provides key information on glacier dynamics, velocity fields also provide useful information on processes such as glacier surges or ice fall, together with reflecting the impact of global warming on movement and surface velocities (Vadon and Berthier, 2004).

However, previous investigation into velocity measurements has been limited due to limited in-situ measurements, and even remotely sensed data has been limited due to issues of severe topography and lack of suitable data (Luckman et al., 2007). More recently microwave remotely sensed data has become available, and provides an advantage of all-weather imagery (removing the problem of cloud cover which is common feature). This has enabled the application of satellite radar interferometry (SRI) which utilises SAR, and measures velocity from displacements between two SAR images (details in section 2.5.3) (Luckman et al., 2007). However, additional DEM data is required, whereas, when using ASTER data, the DEM can be generated from the visible channels (as long as the DEM generated is of suitable accuracy).

ASTER DEM data is now also globally available removing the limitation of a lack of DEM data for many locations. Through the use of multiple ASTER images and ASTER DEMs, the surface
velocities of a debris-covered glacier can be estimated, with feature tracking used as the principal method, and the spatial displacement of features in repeated optical imagery measured (Kaab, 2005). The DEM data are used to generate orthoimages from the ASTER visible data so that the resulting displacements can be directly transformed into horizontal movement, with accuracies around +/-15 m (Kaab et al., 2003).

2.7. Lithology mapping using remote sensing

A number of studies have used remote sensing for the application of rock mapping (e.g. Longhi et al., 2001; Rowan and Mars, 2003; Ninomiya, et al., 2005; Gad and Kusky, 2007; Krezhova et al., 2007). However, this technique has not been previously applied to debris-covered glaciers to map the rock types present on the surface. Information on rock types can be used for the refinement of melt rate calculations, because rock properties, including its albedo and thermal conductivity, vary between different rock types (Table 2.4a and 2.1) and result in different levels of ablation. It could also enable the identification of emissivity at different locations on the glacier (as it varies with rock type), which has implications regarding the estimation of surface temperatures from satellite imagery.

Whilst not currently used to estimate lithology of surface debris, it is important to review the application of remote sensing to estimate lithology. Since Hunt (1989), a number of studies have generated vast spectral libraries on the laboratory spectra of rocks and minerals. This established the basis for the scientific background for the interpretation of remotely sensed data. These spectral libraries highlight the fact that different rocks or minerals can be identified through their spectral signatures, due to the presence of absorption or reflectance features which are in the same locations for similar rock types (Figure 2.6). These features result from a number of factors including the principal mineralogy present within the rocks, with similar rock
types containing similar minerals and therefore displaying similar spectral signatures (Longhi et al., 2001).

Many previous studies applied band ratios to images to emphasize the spectral characteristics of certain rocks and minerals, and display the spectral contrast of specific absorption features (Rowan and Mars, 2003; Gad and Kusky, 2007). Therefore, different rock types stand out from one another, making this method more effective in lithological mapping than using RGB images alone (Gad and Kusky, 2007). A number of different satellite sensors have been used for lithological mapping, including Landsat TM (e.g. Ferrairi et al., 1996), SPOT (e.g. Moghtaderi et al., 2007), and more recently ASTER (e.g. Rowan and Mars, 2003; Ninomiya, et al., 2005; Gad and Kusky, 2007). ASTER’s combination of both a wide spectral coverage and a high spatial resolution in all three of the VNIR, SWIR, and TIR enhances its capability of discriminating different rock types (Gad and Kusky, 2007). Therefore, it provides a better alternative to other satellites including Landsat TM.

Figure 2.6: Reflectance curves for three rock families showing spectral similarities of reflectance and absorption features, a) limestones, b) sandstones, c) shales (Prost, 1994).
Other studies have utilized spectral data. Spectral data can either be obtained from a previously existing database of the spectra of different rock/minerals obtained using a Spectrometer (e.g. USGS spectral library), or samples can be obtained in the field and spectral signatures for those specific samples can be identified using the spectrometer in the laboratory (Longhi et al., 2001). From these signatures, certain spectral features can be identified. These can then be compared to signatures obtained from satellite sensors (e.g. ASTER, which have a smaller spectral range), meaning dominant mineral assemblages can be identified and classified within an area.

### 2.8. Debris thickness estimations from satellite imagery

The estimation of debris thickness on a debris-covered glacier is important in the determination of sub-debris ablation, its distribution, and resulting impact upon water resources from glacial melt in remote locations (Mihalcea et al., 2008). Previous studies have applied empirical approaches based on the relationship between debris thickness and surface temperature, but these are glacier specific and have limited transferability (e.g. Mihalcea et al., 2008). An energy balance approach which utilises remote sensing has potential to resolve this issue, and the components of an energy balance model and its application on debris-covered glaciers will be discussed.

#### 2.8.1. Surface energy balance

The surface energy balance of a glacier considers the fluxes of energy into, out of, and within a terrestrial surface layer (Equation 2.7). It can be used to estimate the amount of melt which occurs on a glacier surface as a result of energy fluxes between the atmosphere and the glacier (Greuell and Genthon, 2004). First, it computes the energy fluxes between the atmosphere and the glacier as a function of meteorological variables and the state of the surface. Second, the state of the subsurface, which is forced by energy exchange with the atmosphere, is considered (Greuell and Genthon, 2004). With terms: $SWR\downarrow$ (incoming shortwave radiation, W m$^{-2}$), $LWR\downarrow$
(incoming longwave radiation, W m\(^{-2}\)), \(LWR\uparrow\) (outgoing longwave radiation, W m\(^{-2}\)), \(SHF\) (sensible heat flux, W m\(^{-2}\)), \(LHF\) (latent heat flux, W m\(^{-2}\)), \(PRE\) (energy from precipitation):

\[
SWR\downarrow + LWR\downarrow - LWR\uparrow + SHF + LHF + PRE
\]  

(2.7)

The presence of a debris layer on a glacier has a considerable impact upon the surface energy balance and is a poorly researched topic (Brock, et al., 2007). Increased research is required into the application of physically-based ablation models designed specifically for debris-covered glaciers (instead of degree day factor models) which can be used to predict both short and long-term responses of melt as a result of differing meteorological conditions (Nicholson and Benn, 2006). A different energy balance equation (Equation 2.8) is required on debris-covered glaciers due to the impact of a debris layer upon an ice surface, through its influence on ablation rates. Resulting in the inclusion \(\partial STOR\) to account for heat stored within a debris layer, along with the reduction or removal of the influence of latent heat fluxes at the surface (Equation 2.8). Where: \(\partial STOR\) defines the change in heat store, and \(LHF\) is assumed to be zero.

\[
SWR + LWR + SHF + LHF + COND + PRE + \partial STOR = 0
\]  

(2.8)

The component of \(LHF\) is often removed because, except during rainfall events, a debris surface is dry and water does not condense upon the surface (although evaporation and condensation will occur within debris layers). Previous attempts at surface energy balance modelling on debris-covered glaciers are provided by Nakawo and Young (1981) and more recently by Nicholson and Benn (2006). A physical model was developed as an alternative to using empirical methods which are site specific (e.g. Mihalcea et al., 2008) and, therefore, cannot be applied for predicting global glacier runoff changes as a result of climate change (Nicholson and Benn, 2006).
2.8.2. Short wave radiation (SWR)

Shortwave radiation is any radiation which is received by the surface that has a wavelength between 0.3-3.0 μm, almost all of which is due to solar irradiance (Oke, 1987). The amount of this radiation which reaches the surface can be measured using a pyranometer. Once measured, SWR can be recalculated for differing slopes and aspects. A variety of methods are also available to empirically determine SWR if measurements are not available. An example of an empirically based approach can be found in Strasser et al., (2004) on Haut Glacier d’Arolla, Switzerland (Equation 2.9-2.13). Where: \( SWR_{\text{dir}} \) (direct component of incoming shortwave radiation, \( \text{Wm}^{-2} \)), \( SWR_{\text{dif}} \) (diffuse component of incoming shortwave radiation, \( \text{Wm}^{-2} \)), \( \alpha \) (Surface albedo), \( SWR_{f} \) (diffuse fraction of total incoming shortwave radiation, \( \text{Wm}^{-2} \)), \( SWR' \) (incoming shortwave radiation measured in a horizontal plane, \( \text{Wm}^{-2} \)), \( Z' \) (angle of slope from the horizontal, degrees), \( Z \) (angle of sun above horizon, degrees), \( SA \) (Solar azimuth, degrees), \( A' \) (aspect of slope, degrees), \( SWR' \) (equivalent radiation received by a surface, \( \text{Wm}^{-2} \)), \( \alpha_m \) (mean albedo of surface).

\[
SWR = (1 - \alpha) SWR_{\text{dir}} + (1 - \alpha) SWR_{\text{dif}} \quad (2.9)
\]

With:

\[
SWR_{\text{dif}} = SWR_{f} SWR' \cos (Z'/2) + \alpha_m SWR' \sin (Z'/2) \quad (2.10)
\]

\[
SWR_{\text{dir}} = (1 - SWR_{f}) SWR' \left( \sin Z \cos Z + \cos Z \sin Z \cos (SA - A') \right) \quad (2.11)
\]

\[
SWR_{f} = (0.65 +0.15) \quad (2.12)
\]

\[
SWR' = SWR' / \sin Z \quad (2.13)
\]
2.8.3. **Long wave radiation (LWR)**

Longwave radiation is radiation received by the surface at wavelengths greater than 3.0 \( \mu \text{m} \). The amount of this radiation which arrives at or leaves the surface can be measured using a pyrgeometer located at a weather station. \( LWR \) can also be determined empirically from the temperature and emissivity of the emitting body, and a number of different methods are available (e.g. Brunt, 1932; Swinbank, 1963; Montieth, 1973). These empirical methods have a number of limitations, which occur because some are limited to specific temperature ranges, with Monteith's (1973) applicable to a range of -5 to 25 \( ^\circ \text{C} \), and Swinbank’s (1963) limited to temperatures above 0\( ^\circ \text{C} \). Despite these limitations, many methods are applicable to ‘clean’ glaciers including that used by Brock and Arnold (2000) (Equation 2.14-2.16). Where: \( k_c \) (constant depending upon cloud type, with a mean value of 0.26 for altostratus, altocumulus, stratocumulus, stratus, and cumulus), \( \varepsilon^* \) (effective emissivity of the sky), \( n \) (cloud cover, ranging 0.0-1.0), and \( \varepsilon_0 \) (being clear sky emissivity), \( T_a (K) \) (Absolute air temperature, K), and \( \sigma \) (Stefan-Boltzmann constant, \( 5.6697 \times 10^{-8} \text{ Wm}^{-2} \text{ K}^{-4} \)):

\[
LWR = (LWR \downarrow - LWR \uparrow) \quad (2.14)
\]

*With:*

\[
LWR \downarrow = \varepsilon^* \sigma T_a (K)^4 \quad (2.15)
\]

\[
\varepsilon^* = (1 + k_c n) \varepsilon_0 \quad (2.16)
\]

2.8.4. **Sensible (SHF) and latent heat fluxes (LHF)**

The sensible heat flux is concerned with the convection of warm air to (when the surface is cooler than the air above it) or from the surface (when the surface is warmer than the air above it), with a portion of the daytime radiative heat surplus being carried into the atmosphere as
sensible heat (Oke, 1987; Paterson, 2001). Its magnitude is proportional to the temperature gradient between the surface and the air above it, the wind speed above the surface, the surface roughness, and the thermo-dynamic structure of the atmosphere (stability/instability).

The latent heat flux is the gains or losses of heat due to evaporation or condensation of water, with heat gained when water in the air above the surface condenses upon the surface, whereas heat is lost when moisture evaporates from it. The amount of heat transferred is dependent upon the amount of turbulence in the atmosphere (i.e. wind speed), with heat transfer increasing with an increase of turbulence (Paterson, 2001). The process of both latent and sensible heat transfer on ‘clean’ glacier surfaces generate significant amounts of melt energy, and, therefore, play an important role in the amount of ablation on a glaciers surface (Brock, et al., 2006). However, studies into the magnitude of turbulent fluxes on debris-covered glaciers are limited (Brock et al., 2007).

A number of approaches have been used to estimate the sensible and latent heat fluxes on a bare ice surface and two are outlined below.

### 2.8.4.1. **Eddy covariance method**

The eddy covariance method directly measures the vertical fluxes in the surface boundary layer, by sensing the properties of eddies as they pass through a measurement height (Oke, 1987). It is based on the fact that all atmospheric entities have short period fluctuations around their long term mean, which occur as a result of turbulence which force eddies to move around, whilst taking their properties with them to another location (Oke, 1987). These eddies consist of a number of properties including density, vertical velocity and the volumetric content of any entity it possesses. The vertical flux of the entity (the momentum, through heat or moisture) can be calculated once these values are known. From this, the sensible heat flux can be calculated
(Oke, 1987). This method requires precise equipment to measure the vertical movement and velocity of the air and, therefore, can be very inaccurate if proper care is not taken with both the sensor design and deployment in the field, particularly its height above the surface. Also, extreme weather conditions experienced on many glaciers may damage the equipment with resulting measurements being incorrect or impossible to obtain (Munro, 1989).

2.8.4.2. Profile methods

The use of profile methods and especially the bulk transfer method are preferred over eddy covariance methods on snow/ice surfaces, due to the required instrumentation being simple to setup and record, enabling long periods of monitoring which result in large data sets in all weather conditions (Munro, 1989). Profile methods aim to infer the vertical flux in the surface boundary layer through the use of average profiles of atmospheric properties and the amount of turbulent activity (Oke, 1987).

Physically-based distribution models/bulk exchange methods

In a simplification of the profile method the bulk aerodynamic method only requires measurements of wind speed (u), humidity, and surface temperature at the surface and one height above the surface, to calculate SHF (Equation 2.17) (Munro, 1989). However, when a transfer coefficient (D) is used, stability within the surface layer is not considered. Where: \( T_a \) (air temperature, \(^\circ\text{C}\)), \( T_s \) (Surface temperature, \(^\circ\text{C}\)), u (wind speed, m\(^s\text{-1}\));

\[
SHF = D \, u \,(T_a-T_s) \tag{2.17}
\]

N.B. \( SHF \) is positive when heat is transferred down from the air to the surface.

The method can be extended to non-neutral conditions (so it considers stability or instability in the surface layer) using the Richardson number (\( R\delta \)) to calculate the transfer coefficient
This number is a means of categorising atmospheric stability (with the state of turbulence) in the lowest layers of the surface boundary layer (Oke, 1987). In stable conditions $R_b$ is positive, whilst in neutral conditions reaches 0. Whether this number is positive or negative determines whether the stable or unstable case equation (Equation 2.19a and b) needs to be applied for the calculation of $D$, and identifies the flow regime at the time of interest (Figure 2.7).

**Figure 2.7:** *Stability functions against the Richardson number, showing the flow regimes that occur at different Richardson number values* (Oke, 1987).

The use of the Richardson number approach extends the usefulness of the basic aerodynamic approach to the application over a range of stability regimes. The aerodynamic roughness length ($z_o$) is the height above a surface at which the extrapolated horizontal wind-speed profile reaches zero (Brock *et al.*, 2006), and is a function of the size, shape, and density distribution of surface roughness elements (Table 2.7). Where: $z_o$ (aerodynamic roughness length, mm), $z_a$ (measurement height, m), $k$ (Von Karmans constant), $\rho_a$ (Air density, kg m$^{-3}$), $c_a$ (specific heat of air, 1010 J Kg$^{-1}$ K$^{-1}$):
\[ R_b = \frac{2g(T_a - T_s)z_o}{(T_a + T_s + 546.4)u^2} \]  

(2.18)

**a)** \( R_b \) Positive (stable case)  

\[ D = (1 - 5R_b)^2 k^2 \rho_o c_a \]  

\[ ((\ln(z_o/z_0))^2 \]  

(2.19a)

**b)** \( R_b \) Negative (unstable case)  

\[ D = (1 - 16R_b)^{0.75} k^2 \rho_o c_a \]  

\[ ((\ln(z_o/z_0))^2 \]  

(2.19b)

---

**Table 2.7: Aerodynamic roughness lengths on both clean and debris-covered glaciers**

<table>
<thead>
<tr>
<th>( z_o ) mm</th>
<th>Surface type</th>
<th>Source</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Tephra covered</td>
<td>Brock et al., (2007)</td>
</tr>
<tr>
<td>16</td>
<td>Debris covered</td>
<td>Brock et al., (in press)</td>
</tr>
<tr>
<td>1</td>
<td>Debris covered</td>
<td>Nicholson and Benn (2006)</td>
</tr>
<tr>
<td>6.3</td>
<td>Debris covered</td>
<td>Takeuchi et al., (2000)</td>
</tr>
<tr>
<td>0.1-5.8</td>
<td>Glacier ice</td>
<td>Brock et al., (2006)</td>
</tr>
<tr>
<td>3-15</td>
<td>Rough glacier ice</td>
<td>Brock et al.,(2006)</td>
</tr>
<tr>
<td>20-80</td>
<td>Very rough glacier ice</td>
<td>Brock et al.,(2006)</td>
</tr>
</tbody>
</table>

The aerodynamic roughness length is an important control upon the rates of turbulent (sensible and latent) heat transfer between a glacier surface and the air directly above it (Paterson, 2001; Greuell and Genthon, 2004; Brock et al., 2006), with an order of magnitude increase in \( z_o \) resulting in a doubling of the turbulent fluxes (Brock, et al., 2000). It indirectly influences, therefore, the melt energy at the surface and, in turn, means that variations of \( z_o \) need to be considered when calculating glacier surface melt rates. The value of \( z_o \) varies with wind direction in the presence of regularly shaped obstacles, meaning a different value of \( z_o \) will be required for a debris-covered glacier which will have a much rougher or irregular surface than clean ice (Table 2.7). Greater spatial variation in \( z_o \) was found by Brock et al., (2006) on the
Haut Glacier, d’Arolla when debris cover was present, due to variability in debris characteristics (shape, size, orientation) at different locations on the glacier.

The bulk aerodynamic approach with Richardson stability correction has previously had limited application to debris-covered glaciers, but Brock et al., (2006) on the Villarrica Volcano, Chile, demonstrated its potential. Where $SHF$ underestimates reached only 25% (as long as a representative $z_o$ value is used), compared to the bulk aerodynamic approach without Richardson stability correction (assuming a neutral atmosphere), which underestimated sensible heat fluxes by up to 55%. An alternative to the simplified Richardson number approach for estimating the heat exchange in a turbulent flow is the Monin-Obukhov similarity theory, details of which can be found in Munro (1989, 1990), Brock and Arnold (2000), and Brock et al., (2006).

**Energy from precipitation (PRE)**

Energy from precipitation considers the heat flux supplied to the debris surface by rain, and resulting energy exchange. This needs to be considered because heat can be transferred to or from a glacier surface, at rates depending upon the temperature difference between the surface and the rainfall which is incident upon it (Greuell and Genthon, 2004). $PRE$ can be calculated if certain variables are known (Equation 2.20). Where: $\rho_w$ (density of water, 1000 kg/m$^3$), $r$ (rainfall rate, mm hr$^{-1}$), $C_{pw}$ (specific heat capacity of water, 4200 J kg$^{-1}$ deg$^{-1}$), $T_r$ (Temperature of rainfall °C):

$$PRE = \rho_w \ r \ c_{pw} \ (T_r - T_s) \quad (2.20)$$

**Change in heat store (\(\partial STOR\))**

$\partial STOR$ (W m$^{-2}$) is an energy flux measuring the rate of change of heat stored in a debris layer. $\partial STOR$ experiences a diurnal cycle due to the debris layer being warmed during the daytime by
$SWR_{\downarrow}$ which increases the amount of heat stored within the debris layer, followed by a decrease in stored heat in the late afternoon/evening as heat from the debris layer is lost to the atmosphere through $SHF$ and $LWR_{\uparrow}$. Hourly variations in $\partial STOR$ are the result of any changes in $SWR_{\downarrow}$ occurring during the day, where changes can result due to cloud blocking out the passage of $SWR_{\downarrow}$ to the ground. Therefore, the change in heat energy stored within the debris layer can be identified over a specific time period (Equation 2.21, Brock et al., 2007). Where: $\rho t$ and $ct$ (density of material, kg m$^{-3}$), $T_i (T_i/2)$, and $\partial T_i$ (average rate of temperature change, K s$^{-1}$):

$$\frac{\partial STOR}{\partial t} = \rho t c_t \frac{\partial T_i}{\partial t} d$$  \hspace{1cm} (2.21)

Brock et al., (2007) calculated $\partial STOR$ at four different sites on Villarrica volcano, Southern Chile, and found net values ranging -0.011 and 0.113 M J m$^{-2}$, which is only equivalent to enough energy to melt 0.03 to 0.3 mm of ice, over a 17 day period. Hence, $\partial STOR$ is only important over sub-daily timescales.

2.8.5. Calculating melt rates (modelling)

Once all of the variables in the surface energy balance equation have been calculated ($SHF$, $LWR$, $SWR$, $COND$), the average amount of melt per hour due to the conductive heat flux can be calculated using the conductive heat flux value, and the latent heat fusion of ice (Equation 2.22). This assumes that all of the heat convected through the debris is used for the melting of the ice and $\partial STOR$ over the calculation period is $\sim$0. Where: $MELT$ (Sub-tephra ice melt rate over unit of time, mm w.e. per hour), and $L_i$ (latent heat fusion of ice, $3.34 \times 10^5$ J Kg$^{-1}$):

$$MELT = \frac{COND}{L_i}$$  \hspace{1cm} (2.22)
2.8.6. Previous application to debris-covered glaciers

A number of studies have applied an energy balance model to debris-covered glaciers, including Nakawo and Young (1981), Kayastha (2000), Nicholson and Benn (2006), and Brock et al., (2007). Each follow a slightly different approach, and the specific features which make up these models will be discussed below.

2.8.6.1. Surface temperature

The calculation of the energy balance model requires surface temperature, with many approaches measuring surface temperature at the study site for application into the model (Brock et al., 2007). However, direct measurements of surface temperature are not always available and two different approaches aim to combat this by either ignoring the surface temperature variable (Nakawo and Young, 1981), or by using an iterative approach (Nicholson and Benn, 2006).

Ignored

Nakawo and Young (1981) proposed that the problem of not having surface temperature measurements can be removed by using meteorological variables, which are easier to obtain than surface temperature values (Nicholson and Benn, 2006), although despite this statement being made, no numerical formulae are provided. However, they do highlight that, since surface temperature contributes to all of the energy balance components’ calculations except $SWR$, as long as all other variables can be measured or estimated surface temperature can be eliminated from the equations, or have an assumed value. For thin debris layer’s the surface temperature of the debris layer can be assumed to be close to the ice temperature.

The energy balance equation applied by Nakawo and Young (1981) (Equation 2.23) is a simplified equation, because, it does not consider changes in stored heat within a debris layer,
and assumes that the temperature gradient between the upper and lower surface of the debris layer is linear. This assumption is based on the notion that the debris layer is in thermal equilibrium, meaning the heat stored within the debris layer is constant over time. Therefore, the conductive heat flux at the debris surface is the same at the ice interface. Where: $G$ (global radiation, W m$^{-2}$), $\beta$ (coefficient of heat transfer, 4.89 J m$^{-3}$ deg$^{-1}$), $L_e$ (Latent heat of evaporation of water, 2.494 x10$^6$ J kg$^{-1}$), $c$ (specific heat capacity of air at constant pressure, 1.0 J kg$^{-1}$ deg$^{-1}$), $e_a$ (vapour pressure in the air, Pa), and $e_s$ (vapour pressure at debris surface, Pa; assumed to be same as $e_a$ on a dry surface, and assumed to be the saturated vapour pressure for the debris surface temperature on wet surfaces):

$$COND = SWR + SHF + LHF$$  \hspace{1cm} (2.23)

With:

$$SWR = (1-\alpha) G + LWR \downarrow - \sigma (T_s + 273)^4$$  \hspace{1cm} (2.24)

$$SHF = \beta u (T_a - T_s)$$  \hspace{1cm} (2.25)

$$LHF = \beta L_e u (0.623/P \times a) (e_a - e_s)$$  \hspace{1cm} (2.26)

**Iterative**

Nicholson and Benn (2006) developed a method which identifies surface temperature through optimization, and eliminating changes in stored heat within a debris layer by using daily means in the simplified energy balance equation (Equation 2.23). Daily means were applied as results from Nakawo and Young (1981) over-predicted melt rates, due to the use of sub-diurnal timescales and changes occurring in the amount of heat stored within the debris during the calculation interval (Figure 2.8).
Figure 2.8: Vertical temperature profiles within a debris layer on Ngozumpa Glacier, Nepal (Nicholson and Benn, 2006).

This highlights that at sub-diurnal timescales the relationship of temperature and depth in a debris layer is not linear, however, on a diurnal timescale (with stable weather conditions) the linear assumption is more likely to occur (Conway and Rasmussen, 2000; Nicholson and Benn, 2006) (Figure 2.8). They also note that if all other variables can be measured or estimated, $SHF$, $LHF$, and $COND$ can be solved using daily mean values of the required meteorological variables. Then, surface temperature can be found by iteration, as an assumption is made that both the surface temperature is the mean daily value and that all of the surface temperature dependent fluxes are linear functions of the surface temperature (Nicholson and Benn, 2006).

**Measured**

Surface temperature can be measured in the field using a number of methods including surface thermistors (e.g. Brock et al., 2007), or an infrared thermometer and photodiode (e.g. Kayastha, et al., 2000). These values can then be placed directly into the energy balance component calculations.
2.8.6.2. Surface layer (in)stability in turbulent fluxes

One limitation of the method used by Nakawo and Young (1981), and Kayastha et al., (2000), is that although they used the bulk aerodynamic approach both the SHF and LHF determination methods use only empirical transfer coefficients which neglect atmospheric (in)stability. Nicholson and Benn (2006) also assumed a neutral atmosphere and, therefore, the impact of variation in atmospheric stability on the turbulent fluxes is not considered. Brock et al., (2007) did take into account the impact of atmospheric (in)stability, by extending the bulk aerodynamic approach to non-neutral conditions by using the bulk Richardson number; a method which previously had not been applied to debris-covered glaciers.

2.8.6.3. Thermal properties of debris layer

Nakawo and Young (1981) assumed that the stored heat in the debris layer is constant, which neglects changes in stored heat which could be significant in thick debris layers with large heat capacities. This problem is most prominent in debris covers greater than 0.5 m thick (Haidong et al., 2006), and results in an overestimation of ablation calculations (Nicholson and Benn, 2006). Nicholson and Benn (2006) utilise the same energy balance equation as Nakawo and Young (1981), and also eliminate changes in stored heat within a debris layer (Equation 2.23), by using a simplified equation (also used by Kayastha et al., 2000, and Brock et al., 2007). However, the use of daily averages removes the problem of ablation overestimates.

Nicholson and Benn (2006) considered two different states in the calculation of both SHF (Equation 2.28) and LHF (Equation 2.27). First, a dry state when a debris layer is dry (no condensation and evaporation) and the influence of latent heat fluxes can be ignored as the flux is zero. Second, a wet state during which the debris surface is wet and therefore latent heat fluxes can operate through condensation and evaporation. However, it assumes that the debris layer is completely saturated, meaning the latent heat flux in partially wet conditions cannot be
determined (Equation 2.28). Where; $P_0$ (standard air pressure at sea level, $1.0 \times 10^5$ Pa), $p_0$ (density of the air at standard sea-level pressure, $1.29$ kg m$^{-3}$), $P$ (air pressure at the site, Pa), $c$ (specific heat capacity of air at constant pressure, $1.0$ J kg$^{-1}$ deg$^{-1}$), and $A$ (dimensionless transfer coefficient):

$$LH \text{F} = (0.622 \, p_0/P_0) \, L \, e \, u \, (e_a - e_s) \quad (2.27)$$

$$SH \text{F} = p_0 \, (P/P_0) \, c \, A \, u \, (T_z - T_s) \quad (2.28)$$

With:

$$A = \frac{k^2}{(\ln (z/z_o))^2} \quad (2.29)$$

One limitation, however, is that this method (in both the wet and dry state) assumes that both the longwave radiation and turbulent heat fluxes are linearly dependent upon surface temperature. This incorrect assumption will probably result in over-prediction of the sub-debris ablation rate because, by using only the daily mean value of temperature, $SHF$ and $LWR$ are underestimated during the daytime when surface temperature is high. Also, the method was only tested with two short periods of data and, therefore, needs to be applied and tested elsewhere for validation.

**Temporal resolution**

The equation utilised by Kayastha et al., (2000), Nicholson and Benn (2006), and Brock et al., (2007) to calculate the COND (Equation 2.4) assumed that the temperature gradient between the upper and lower surfaces of the debris layer is linear. However, looking at vertical thermal profiles of a debris layer over a sub-daily timescale it is possible to see that the profiles are not linear, but when an average is taken of the daily temperatures a linear profile does result (Figure 2.8). As a result, this means that the assumptions made of a linear profile are met on a diurnal
timescale, but are not met on sub-diurnal timescales. Therefore, the thermal conductivity equation would not be suitable for use on anything less than a 24 hour energy balance model. Consequently, Kayastha et al., (2000) and Nicholson and Benn (2006) used it for daily estimates, and Brock et al., (2007) for both daily and ablation season estimates.

### 2.8.6.4. Distributed modelling

All of the methods discussed above were applied at the point scale but extension to a glacier-wide scale requires distributed modelling. Hock and Holmgren (2005) applied a distributed surface energy balance model to estimate mass balance on Storglaciaren, Sweden, by calculating the energy balance Equation (2.12) for every hour at each grid cell (of a 30 m resolution DEM) using meteorological data collected at one point on the glacier. Details of the process applied can be found in Hock and Holmgren (2005) and Reijmer and Hock (2008). Results from these models need to be validated using field measurements, which is an essential process to tie the models to real ground conditions (Machguth et al., 2006).

Application to debris-covered glaciers are limited in number. One example is Mihalcea et al., (2008), where a distributed energy balance model was applied to Baltoro glacier, Karakoram, Pakistan to estimate sub-debris ablation. This study combines both meteorological data and thermal remote sensing data for determining debris thickness through a generated equation using surface temperature from the thermal image and measured debris thicknesses in the field. Consequently, this highlights that remote sensing can make a useful contribution to distributed melt modelling, by producing maps of both debris thickness and extent at a spatial scale which could not be obtained using field data alone.

One of the main limitations of using a distributed model approach is that the spatial distribution of debris thickness across the glacier, its thermal properties, and moisture content (or its
variability), is unlikely to be known in enough detail or with enough accuracy to enable the physically-based calculation of surface temperature using meteorological measurements alone (Brock et al., 2007). Even if surface temperature is obtained from satellite imagery, the spatial scale of the imagery (90 x 90 m for ASTER, 120 m Landsat TM, 60 m Landsat ETM+) is too coarse to provide enough detail of spatial variability in surface temperature (Brock et al. 2007). However, despite the 90 x 90 m resolution of ASTER imagery, Mihalcea et al., (2008) highlight that the sub-debris melt outputs of their study are acceptable in relation to the actual size of the debris covered area of the study site, being 124 km².

2.9. Key research questions

During the completion of the literature review and investigation into previous work carried out on debris-covered glaciers a number of key research questions became apparent, which provide justification to the research aims and objectives identified at the start of this project (section 1.2). These questions, therefore, set the context of this study, and are listed below.

- **Why focus on debris-covered glaciers?**
  The number of studies on debris-covered glaciers is significantly lower than those on ‘clean’ glaciers, in turn, there are a smaller number of transferable techniques for the monitoring of these glaciers. This highlights the requirement for an increased focus on debris-covered glacier monitoring, along with the development of new techniques and testing of previously developed techniques to enable debris-cover glacier monitoring on a global scale.

- **Can a transferable method to estimate debris thickness be developed?**
  The potential to map debris thickness using satellites has been partially demonstrated, however, this has not been fully realised due to the use of empirical (e.g. Mihalcea et al., 2008) rather than
physically-based algorithms. This has, therefore, restricted the development of transferable debris thickness estimation models.

- **Can debris extent variations over time be monitored?**

  Despite the boundaries of debris-covered glaciers being mapped remotely, few studies have looked into monitoring the changes in debris extent over time. Changes of debris cover extent are of key importance, because an increase in debris layer has a considerable impact upon a glaciers mass balance by insulating the ice beneath, and reducing ablation rates (as long as a the debris layer is thick, the opposite will be encountered if a thin layer is present). This also links to mapping debris thickness and any changes which occur over time, because variations in both the extent and thickness of a debris layer will have a significant impact upon surface ablation and resulting run off rates.

- **Can surface elevation changes over time be identified with ASTER DEM imagery?**

  Previous attempts to monitor surface elevation changes using ASTER DEMs have been made, however, their successes have restricted by accuracy errors of the ASTER DEMs applied. Whether these errors occur at all study locations (and to what magnitude) needs to be assessed, as the potential of the ASTER DEM dataset is significant for debris-covered glacier monitoring on a global scale due to its global coverage at a high temporal resolution.

- **Is ASTER visible and DEM data suitable for obtaining surface velocity estimations?**

  The potential of ASTER DEM data to obtain surface velocity measurements has been highlighted, however, few studies have focussed upon this method, and in turn its suitability requires assessment.
• **Can different rock lithologies present within a debris layer be mapped?**

The presence of different rock types will affect the surface ablation rate (due to their different albedo, thermal conductivity and emissivity values). Therefore, a means of identifying the spatial distribution of different rock types is required. However, despite the previous successes of remotely sensed data for mapping rock types in non-glaciated terrain (e.g. Gad and Kusky, 2007; Rowan and Mars, 2003), a means of mapping the rock types present within a debris layer using satellite data has not been previously developed.

### 2.10. Summary

From the literature it is clear that there are a number of current limitations in the monitoring of debris-covered glaciers, including their locations which are often difficult to access, and the presence of extensive debris covers hindering fieldwork on their surface. The size of some debris-covered glaciers also makes it difficult to monitor them on a large scale using field based methods alone. Consequently, fewer studies have been completed on debris-covered glaciers, and have focused mainly on point location sub-debris ablation rates using data collected in the field inputted into energy balance models.

Remote sensing provides a solution to many of these problems, as imagery can be obtained at any location, removing the issue of accessibility. Therefore, the potential for debris-covered glacier monitoring using remote sensing is highlighted, because, previously it has had limited application to debris-covered glaciers focusing upon boundary identification, surface elevation change monitoring, and surface velocity measurements. However, it could also be applied to obtain information on both debris cover thickness and extent, to aid in the identification of ablation rates on a glacier-wide scale using energy balance models.
This chapter has also shown the increasing complexity of the surface processes occurring on debris-covered glaciers compared to ‘clean’ glaciers, and is due to the presence of a debris layer, which increases the complexity in estimating ablation rates from debris-covered glaciers, and results in the requirement of a modified energy balance equation. The presence of this debris layer has protected many glaciers from the current trend of retreat experienced by many ‘clean’ glaciers, highlighting the importance in the monitoring of these debris-covered glaciers and the processes occurring within them and the need for an increased number of techniques to achieve this. The continued development and modification of remote sensing techniques will enable the increased monitoring of these glaciers due to the increased spatial resolution and temporal frequency compared to field based observations and, as a result, can be applied to a larger number of glaciers.

Therefore, this project addresses these issues and increases the possibilities and techniques available for debris-covered glacier monitoring such as the development of a method to estimate debris cover thickness on a glacier-wide scale. By increasing the monitoring techniques greater amounts of information can be obtained from debris-covered glaciers, which in turn will enable an increased understanding of their behaviour and response to changes in climate.
CHAPTER 3: STUDY SITE

3.1. Introduction

This chapter presents details of the study site of the Miage Glacier, including its location and key characteristics. Justification for the selection of the Miage glacier as the study site is also included.

3.2. Site description

The Miage Glacier (Ghiacciaio del Miage) is situated on the Italian side of Mont Blanc (Monte Bianco) in the western Alps of Italy. The Alps are a complex system of mountain ranges in south-central Europe which extend 1,000 km along Italy’s common border with Austria, Switzerland and France in a crescent shaped line (Williams and Ferrigno, 1993). The total surface coverage of all glaciers, ice, and permanent snow in the Italian Alps is 608 km², and comprises approximately 1,400 glaciers (Barbero and Zanon, 1993).

The Miage Glacier (45° 47’ 30’’ N, 6° 52’ 00’’E) is a temperate glacier and is located a few kilometres west of Courmayeur in the Valley of Aosta, with the glacier being morphologically very similar to large Asian debris-covered glaciers (Smiraglia, et al., 2000) (Figure 3.1). It is the most famous debris-covered glacier in the Italian Alps (Pelfini, et al., 2007), and its accumulation basin is made up of three steep (~24-33 °) convergent glacial tributaries which include the Mont Blanc, Dome, and Bionnassay, and one secondary ice mass Tête Carée (Tinti et al., 1999). These four masses converge into the Miage Glacier valley to form the more gently inclined glacier tongue (~5 °) which extends to 5 km in length to the South East (Thomson et al., 2000). Its total length of 11-13 km makes it the largest debris-covered glacier in the Italian Alps (Deline and Orombelli, 2005).
Figure 3.1: Map showing the location of the Miage Glacier within the Mont Blanc Massif – the Western Alps (Deline, 2005).

The glacier is covered in debris below an altitude of 2500 m a.s.l, initially by medial moraines (Diolaiuti et al., 2005), but then becoming continuous extending over 6 km of the glacier tongue in total (Deline and Orombelli, 2005) (Figure 3.2). Above 2500 m a.s.l. debris cover is thin and patchy. Debris is supplied mainly due to rockfalls and avalanche events at exposed headwalls between the tributary glaciers (Thomson et al., 2000; Pelfini, et al., 2007). Where the glacier tongue reaches the east - west aligned Val Veny it is deflected to the left by massive lateral moraines which extend up to 140 m in height. Below this point the tongue splits into two lobes in its lower part (which are much steeper than the previous areas ~11 °). These terminate at
1730 m and 1775 m a.s.l. with both lobes being bounded by moraines (Smiraglia, et al., 2000; Deline and Orombelli, 2005).

**Figure 3.2:** Map of the Miage Glacier showing its physical characteristics (Thomson et al., 2000).

The debris thickness at the terminus of the glacier generally exceeds 0.5 m, this zone is also covered by sparse vegetation consisting of small trees such as fir and larch (Ranzi et al., 2004; Pelfini. et al., 2007), whereas the median thickness over much of the debris-covered area is about 0.2 m (Thompson et al., 2000). The development of the Miage Glacier into a debris-covered glacier has modified its present day behaviour in comparison to other alpine glaciers, which are currently distant from their Little Ice Age (LIA) maximum positions (Pelfini et al., 2007). Mass variations on the Miage Glacier are translated into thickness changes in the glacier which as a result thickens or thins with the passage of kinematic waves down glacier with limited lateral movement at the snout (Pelfini et al., 2007).
3.3. Study site selection

The Miage Glacier was selected for this study for a number of reasons, the main being the archive of previous work available on the glacier. As well as previous studies undertaken on the glacier (e.g. Deline, 1999; 2002; 2005; 2009; Deline and Orombelli, 2005; Mihalcea et al., 2008; Diolaiuti et al., 2009) this included automated weather station (AWS) data, field measurements of debris thickness, ablation rates, surface temperatures and digital elevation models (DEM). In turn, this provided a wealth of existing data which could be utilised (and increased) during this study.

One of the main reasons that such a large amount of data is already available and another reason for the selection of the glacier is its accessibility, which compared to many other glaciers (both debris-covered and ‘clean’) is relatively easy. Consequently, large amounts of equipment can be carried onto the glacier and be set up without too much difficulty, especially on the glacier snout. However, the upper reaches of the glacier provide much more of a challenge compared to the lower reaches, mainly due to the long length of the glacier and steeper upper reaches. Therefore, this explains why a larger amount of data is available for the lower reaches of the glacier.

The size of the glacier is another reason why it was selected, as many other debris-covered glaciers are significantly larger in size, such as Baltoro Glacier, which reaches 62 km in length (Paul et al., 2004) compared to the 13 km of the Miage Glacier. This smaller size is preferable for the development and testing of methods before their application to much larger glaciers, which are more difficult to obtain data from due to their much larger size.
CHAPTER 4: METHODOLOGY

This chapter focuses upon the processes applied and the methods employed to collect field data during this study. Justification for the field data methods utilised is also provided. Along with an introduction to the satellite images and DEMs used and processing steps completed (and why) prior to their application.

4.1. Methodology/research design

During this project the research undertaken was designed to ensure that it would provide answers to the key research questions outlined in section 2.9 and therefore enable the achievement of the aims and objectives of this project. Consequently, a number of steps were completed for each research question. Every time a step was completed an assessment to whether it had achieved its aim was completed and if it did not achieve its aim, then an additional step was added until the aim was fulfilled. The steps completed are outlined below.

- **Can a transferable method to estimate debris thickness be developed?**
  
  1. Identification of a relationship between debris thickness and surface temperature.
  2. Validation of surface temperature values obtained from ASTER AST08 imagery against values obtained in the field.
  3. Development and application of an empirical approach to estimate debris thickness from AST08 imagery, based on the relationship between debris thickness and surface temperature.
  4. Application of an energy balance model to ensure its successful application.
  5. Development of the energy balance model to enable the calculation of debris thickness at a glacier wide scale from AST08 imagery.
  6. Testing of model to ensure successful application against previous debris thickness estimates from an empirical approach of Mihalcea et al. (2008).
7. Improvement of the model to make it more physically-based to improve debris thickness estimations.

8. Application to a different AST08 image to ensure transferability.


- **Can debris extent variations over time be monitored?**
  1. Application of both a manual and semi-automatic approach to estimate debris extent.
  2. Development of methods to address problems encountered (sporadic debris, steep medial moraines).
  3. Precision analysis.
  4. Identification of debris extent changes over time.

- **Can surface elevation changes over time be identified with ASTER DEM imagery?**
  1. Error analysis of ASTER DEMs (Geographical, vertical).
  2. Identification of surface elevation changes.
  3. Investigation into accuracy of surface change estimates.

- **Is ASTER visible and DEM data suitable for obtaining surface velocity estimations?**
  1. Calculation of surface velocities from an ASTER DEM using a feature tracking approach.
  2. Identification of errors within the velocity image and their cause.

- **Can different rock lithologies present within a debris layer be mapped?**
  1. Completion of spectral analysis (using a spectroradiometer) on rock samples collected in the field.
  2. Investigation into whether rock types present on the Miage Glacier are spectrally different, to enable the generation of a map of rock type distribution using remotely sensed data (based on their spectral differences).
  3. Completion of a supervised classification, this is possible due to spectral differences of different rock types.
4. Testing of supervised classification map of rock types against field data.

5. Completion of an unsupervised classification, to see if rock types could be mapped without any field data.

6. Testing of unsupervised classification map of rock types against field data.

### 4.2. Field data collection

Data used in the project were collected during 5 field work campaigns carried out in the ablation season, June – September of 2005, 2006, 2007, 2008, and 2009. This included continued monitoring of the surface temperatures of the debris as well as a number of meteorological observations.

<table>
<thead>
<tr>
<th>Variable</th>
<th>Sensor</th>
<th>Accuracy under laboratory conditions</th>
<th>Range</th>
</tr>
</thead>
<tbody>
<tr>
<td>$T_a$</td>
<td>HMP45C Probe, Vaisala, Finland</td>
<td>±0.2 °C &lt;0.2 °C at 1100 Wm$^{-2}$</td>
<td>-39-60 °C</td>
</tr>
<tr>
<td></td>
<td>Artificially ventilated in 2005 with RM Young Aspiration shields. 2006 and 2007, naturally ventilated with 41003 Gill radiation shield, RM Young</td>
<td>0.4 °C at 3 m/s at 1080 Wm$^{-2}$ 0.7 °C at 2 m/s at 1080 Wm$^{-2}$ 1.5 °C at 1m/s at 1080 Wm$^{-2}$</td>
<td></td>
</tr>
<tr>
<td>$T_s$</td>
<td>CRN1 Net Radiometer</td>
<td>± 2K (non stable conditions)</td>
<td>-40-+80 °C</td>
</tr>
<tr>
<td>$u$, and wind direction</td>
<td>Tiny Tag data logger with standard 10K NTC (negative temperature coefficient) thermistor probe, Gemini data loggers, UK</td>
<td>Accuracy 0.2 °C at 0 °C</td>
<td>-30-+50 °C</td>
</tr>
<tr>
<td></td>
<td>A100L2 low power Anemometer, Campbell scientific, USA (lower station in 2005, upper station in 2006-7). RM Young, wind monitor 05103-L (lower station in 2005)</td>
<td>1% ±0.1 ms$^{-1}$ Speed: ±0.3 m/s Direction: ±3 °</td>
<td>0-75 ms$^{-1}$</td>
</tr>
<tr>
<td>RH/AVP</td>
<td>HMP45C Probe, Vaisala, Finland</td>
<td>±2 % 0.1-90 % RH; ±3 % over 90-100 % RH</td>
<td>-40 - +60 °C</td>
</tr>
<tr>
<td>$SWR\downarrow$, $SWR\uparrow$</td>
<td>CRN1 Net Radiometer: CM3 Pyranometer, Kipp and Zonen, The Netherlands</td>
<td>Non linearity &lt;±2.5 % $T_a$ dependency &lt;± 6 % Tilt error &lt;±2 %</td>
<td>305-2800 nm</td>
</tr>
<tr>
<td>$LWR\downarrow$, $LWR\uparrow$</td>
<td>CRN1 Net Radiometer: CG3 Pyrgeometer, Kipp and Zonen, The Netherlands</td>
<td>Non linearity &lt;±2.5 % $T_s$ dependency &lt;±6 % Tilt error max 3 %</td>
<td>5-42 μm</td>
</tr>
</tbody>
</table>
Table 4.2: Meteorological station details, see Figure 4.9 for locations

<table>
<thead>
<tr>
<th>Year</th>
<th>Variables measured</th>
<th>Periods of data</th>
<th>Total days</th>
</tr>
</thead>
<tbody>
<tr>
<td>2005</td>
<td>Lower weather station (LWS): $T_a$, $u$, $AVP$, $RH$ (at 0.65, 1.28, and 2.16 m above debris surface), $SWR_{↓}$, $SWR_{↑}$, $LWR_{↓}$, $LWR_{↑}$ (CRN1 at 1.72 m and pyranometer 1.92 m above the surface), $T_s$, U and wind direction (measured at 2.45 m)</td>
<td>15/06 – 08/09</td>
<td>94</td>
</tr>
<tr>
<td></td>
<td>All variables sampled at 10 minute intervals</td>
<td>06/06 – 08/09</td>
<td>85</td>
</tr>
<tr>
<td></td>
<td>Upper weather station (UWS): $SWR_{↓}$, $SWR_{↑}$, $RH$, $AVP$, $u$ (all instruments at 2metres), $T_s$</td>
<td>05/06 – 05/09</td>
<td>92</td>
</tr>
<tr>
<td></td>
<td>Lower weather station: $T_s$</td>
<td>05/06 – 07/09</td>
<td>94</td>
</tr>
<tr>
<td></td>
<td>$SWR_{↓}$, $SWR_{↑}$, $LWR_{↓}$, $LWR_{↑}$ (all at 1.40 m)</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>$RH$, $AVP$, $u$ and wind direction, $T_s$ (all at 2.06 m)</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>All variables sampled at hourly intervals</td>
<td></td>
<td></td>
</tr>
<tr>
<td>2006</td>
<td>Upper weather station (UWS): $SWR_{↓}$, $T_a$, $RH$, $AVP$, $u$ (mean, and max) (all instruments at 2 m), $T_s$</td>
<td>22/06 – 02/09</td>
<td>72</td>
</tr>
<tr>
<td></td>
<td>Lower weather station: $T_a$</td>
<td>19/06 – 05/09</td>
<td>78</td>
</tr>
<tr>
<td></td>
<td>$SWR_{↓}$, $SWR_{↑}$, $LWR_{↓}$, $LWR_{↑}$ (all at 1.7 m)</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>$RH$, $AVP$, $u$ and wind direction, $T_a$ (all at 2.06 m)</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>All variables sampled at hourly intervals</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

The 2005 dataset is an extensive dataset used throughout this thesis. However, it was collected prior to the start of this study during a joint University of Dundee and University of Milan research project. The field research infrastructure established for meteorological monitoring and (debris) surface sampling used throughout this thesis is outlined in Tables (4.1 and 4.2).

4.2.1. Meteorological data – Automatic Weather Station (AWS)

Meteorological data were collected for three months on each fieldtrip using Automatic Weather Stations (AWS) (Figure 4.1). In 2005, AWS were located at one site on the lower glacier (LWS) (location UTM 0332597 5072069, 2030 m), whereas in 2006 and 2007 an AWS was deployed at both the LWS site and at a higher location close to the upper limit of continuous debris cover (UWS) (UTM 0332635 5074230 2340 m) (Figure 4.2).
4.2.1.1. Longwave radiation (LWR) and Shortwave radiation (SWR)

Incoming and outgoing LWR fluxes were evaluated by correcting for LWR emitted by the CRN1 instrument using measurement of the sensor temperature, according to the manufacturer’s instructions. No corrections were applied to the SWR data.
4.2.1.2. Air temperature data (HMP45C sensors)

Air temperature was measured using HMP45C sensors which were shielded to protect the sensors from direct and reflected solar radiation which would affect the data collected. In 2005, the sensor was artificially ventilated whereas in 2006 and 2007 it was naturally ventilated. Due to a problem with the temperature sensor fans in 2005, some of the morning temperature data had to be corrected as the fans were not working during this time, which meant the temperature sensors themselves heated up, resulting in the recorded temperatures being inaccurate. This problem did not occur on all days but only on those when solar radiation was strong and when the battery was not charged enough for the ventilator to start working straight away.

![Correction applied to air temperature data; example of 01/08/05.](image)

Figure 4.3: Correction applied to air temperature data; example of 01/08/05.

The ventilating fans started working at 09:50, which meant data 1-2 hours before this had to be corrected due to the sun heating up the ground, air and sensors themselves before this time. To correct this problem a method of interpolation was used whereby the data at the time when the fans were not working were smoothed to fit in with the data before and after the affected period so that the sudden high peak in temperatures was removed (Figure 4.3). Removal could be
completed, as a start and end point to the problem was known, and the recorded battery power could be used to identify when the fans were not working. \( SWR_\downarrow \) was used to calculate when the sun started heating up the sensor as well as the surrounding air and, therefore, when it needed to be ventilated.

### 4.2.2. Surface temperature data

Measurements of surface temperature were obtained at different locations on the glacier surface during the field work campaigns using thermistors which consisted of 1 or 2 sensor probes.

When putting these thermistors in place, each thermistor probe was taped to the top of a flat rock, with the last few centimetres of the thermistor probe exposed as this is the most sensitive part, ensuring that the tip was in direct contact with the rock beneath it. This rock was then positioned at the sample point (buried depth 0.01 – 0.03 m) and left to record.

<table>
<thead>
<tr>
<th>Year</th>
<th>Thermistor positioning</th>
<th>Duration</th>
</tr>
</thead>
<tbody>
<tr>
<td>2005</td>
<td>One probe taped to top of the rock, second probe taped to the underside</td>
<td>(~ 97 \text{ days (15/06/05 – 20/09/05)})</td>
</tr>
<tr>
<td>2007</td>
<td>One probe taped to top of the rock, second probe at the same location but on: * Different rock type (Figure 3.11 b) * Shadow (Figure 3.11 c) * Debris thickness thinner than the other thermistor (Figure 3.11 d)</td>
<td>(~ 76 \text{ days (20/06/07 – 05/10/07)})</td>
</tr>
</tbody>
</table>

Details of the duration of recording in 2005 and 2007 (along with sensor probe positioning) are found in Table 4.3. Variation in second probe location in 2007 was completed so that the influence of surface microtopography on surface temperature measurement could be investigated.
Figure 4.4: Location of two thermistor probes at four site locations a) 2007 with one sensor on the top of the rock and the other on the bottom, b) 2007 one sensor on a lighter granite and another on a darker schist, c) one sensor in slight shade one in direct sun, d) one on thicker and the other on thinner debris layer near exposed ice.

Figure 4.5: Surface temperature and stake locations in 2005 (ASTER 2000, VNIR1).
The thermistors were located in different positions for data collection. In 2005, the thermistors were located at 25 stake points on the glacier surface at the same locations as ablation stakes. These were located through the glacier, with variable distances between the thermistors ensuring complete glacier coverage in the upper, middle and lower reaches (Figure 4.5). In 2007, thermistors were set out in transects to test the spatial variability of surface temperature. Two transects in June were set out at 2 m intervals for 2 Days (20/06/07 to 21/06/07) (Figure 4.6), at 5 m intervals for 68 days (22/06/07 to 29/08/07) (Figure 4.6), and one transect in August (30/08/07 to 05/10/07) at 20 m intervals for 6 days (Figure 4.6).

Figure 4.6: Location of surface thermistors during 2007 field campaign (ASTER 2000, VNIR 1).

4.2.3. Debris thickness measurements

Debris thickness was measured by digging through the debris layer to the ice beneath. Next, the thickness from the debris surface to the ice was recorded. Locations of debris thickness
measurements are outlined in Table 4.4 and Figure 4.7. To obtain debris thicknesses over an area representative of an ASTER pixel transects were utilised, where debris thickness was measured at equal intervals along a line in a north, south, east, and west direction. From this an average value of debris thickness for the sample area could be calculated, which provided good comparison for the ASTER debris thickness estimates. This is because, the ASTER debris thickness estimates were obtained from a surface temperature value, which was an average of a 90 x 90 m area.

Transects also were most suited to this due to the time required to obtain debris thickness measurements. If measurements had been taken at regular intervals throughout an area (such as every 10 m²) representative of an ASTER pixel (90 x 90 m) a significant amount of time would be required to collect the data, since it took at least a day to obtain a transect of data with ~40 thickness measurements. Transect lengths in the upper reaches in 2006 and 2007 varied due to bare ice areas, ice cliffs, and steep moraine slopes which prevented safe access.

These transects were targeted so that the spatial variability of debris thickness over small and larger spatial scales could be investigated and, in turn, the applicability and representativeness of the 90 x 90 m spatial resolution of the ASTER resolution for identifying debris thickness could be analysed. The aim of each transect was for it to be located with the mid point of the transect located in the middle of a pixel. However, one limitation of the transect data was that each transect was not taken within a single ASTER pixel, due to problems of matching pixel locations on an image to the same location in the field. However, this would also be a problem for samples taken at regular intervals throughout an area the size of an ASTER pixel (and any other sampling strategy) without very accurate GPS equipment, highlighting the difficulties of obtaining validation data in the field for remote sensing studies.
Figure 4.7: Debris thickness measurement locations in 2005, 2006, and 2007, with zoomed in detail of 2007 transect in the upper reaches of the glacier (ASTER 2000, VNIR 1).

Table 4.4: Location of debris thickness measurements in 2005, 2006 and 2007

<table>
<thead>
<tr>
<th>Year</th>
<th>Transect location</th>
<th>Sampling interval</th>
</tr>
</thead>
<tbody>
<tr>
<td>2005</td>
<td>Measurements taken at each of the 25 stake location points (Figure 3.12)</td>
<td>Randomly located</td>
</tr>
<tr>
<td>2006</td>
<td>Lower reaches (04/06/05)</td>
<td>10 m sampling over 100 m in a north, south, east, west direction</td>
</tr>
<tr>
<td></td>
<td>Middle reaches (08/06/05)</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Upper reaches (06/06/05)</td>
<td>10 m sampling over 100 m in a north west direction</td>
</tr>
<tr>
<td>2007</td>
<td>Lower reaches</td>
<td>20 m sampling over 100 m in a south and west direction</td>
</tr>
<tr>
<td></td>
<td>Upper reaches</td>
<td>5m sampling in a north (15 m length), south (10 m length), east (15 m length), and west (25 m length) direction</td>
</tr>
</tbody>
</table>

Single point data was least suited for debris thickness measurements as it did not provide data over and area of similar size to an ASTER pixel, and therefore, was not suited for debris thickness estimations from ASTER imagery. However, this was the only thickness data available for 2005.
4.2.4. Ablation measurements

Estimates of ablation were made between June and September 2005, at 25 locations which corresponded to the locations of the 2005 surface thermistors (Figure 4.5). Holes were drilled into the ice using a hand drill, and 3 m stakes made from a low conductivity white hollow plastic tube were inserted (to minimise any melting which may result due to the conduction of heat through the stake into the ice). Stakes were inserted into the ice between 12-20\textsuperscript{th} June 2005 and the height of the stake to the ice and debris layer was measured, it was measured again a month later at dates between 7-27\textsuperscript{th} July 2005. They were next measured in August and finally in September between 7-26\textsuperscript{th} (Brock \textit{et al.}, in press). During measurement care was taken to ensure minimum disturbance to the debris layer so ablation rates would not be altered during the course of measurement due to human induced debris movements at the surface. Ablation was calculated by subtracting the stake height above the ice in one month from another it (e.g. subtracting July from June, Equation 3.1). Where: \( Ab \) (ablation, m), \( Sh1 \) (stake height 1, m), \( Sh2 \) (stake height 2, m)

\[
Ab = Sh1 - Sh2 \tag{3.1}
\]

4.2.5. Relationship of air temperature to surface temperature measurements

During the 2007 field work campaign both air temperature and surface temperature were measured simultaneously using a portable AWS consisting of a surface thermistor and a ventilated and shielded HMP45AC (Figure 4.8).
These measurements were made along two transects on the 31/08/07. The first was taken in a southerly direction, with data recorded every 20 m for 110 m. On the second westerly transect, only one data measurement was recorded due to the battery cable coming loose and failing to
record data at the other locations over a 100 m transect (with 20 m intervals). Recordings were also collected at 10 locations on the 03/09/07 so that a study of the variability of surface temperature at different elevations could be undertaken (Figure 4.9). At each sample point, data were recorded for a total of 5 minutes, which meant that the first 2 minutes of measurement would enable both the sensors to acclimatise to the surrounding temperature at that location. Therefore, these data were excluded from subsequent analysis. The second 2 minute recording was assumed reliable and used in further analysis.

### 4.2.6. Rock lithologies

Details on the lithology of the debris cover were collected in 2007 and 2009. In 2007, rock samples were collected at 17 sites across the glacier (Figure 4.10). A field sketch was also generated, identifying the distribution of dominant rock types on the glacier (Figure 4.11). Rock types were also recorded at 9 locations in 2009 (Figure 4.12), with the most dominant rock type identified along with any other rock types which were also present.

![Figure 4.10: Location of rock sample collection points.](image)
Figure 4.11: Field sketch showing rock type distribution in 2007.

Figure 4.12: Rock lithology sample points 2009.
4.3. Remotely sensed data and image pre-processing

A variety of remotely sensed images were used in this project (Table 4.5). These were from Landsat (MSS, TM and ETM+) and the TERRA ASTER sensor (Table 4.6). The images acquired by both the ASTER and Landsat sensors had to undergo a number of correction procedures before they could be used.

Table 4.5: Satellite data used (1 Aster-web NASA, 2008; 2 Lang and Welch, 1999; 3 Land Process Distributed Active Archive Centre, 2008)

<table>
<thead>
<tr>
<th>Satellite/Sensor</th>
<th>Image Type</th>
<th>Resolution (m)</th>
<th>Date(s)</th>
<th>Level of pre-processing</th>
<th>Accuracy</th>
</tr>
</thead>
<tbody>
<tr>
<td>TERRA ASTER</td>
<td>Surface kinetic temperature image (AST08)</td>
<td>90</td>
<td>01/08/05, 29/07/04</td>
<td>Level 2 - Temperature-emissivity separation algorithm applied to atmospherically, radiometrically, and geometrically correct surface radiance data</td>
<td>Absolute: 1-4 K, Relative: 0.3K</td>
</tr>
<tr>
<td></td>
<td>Orthorectified images – all channels, and DEM (AST14DMO)</td>
<td>DEM – 30, Ortho: VNIR: 15, SWIR: 30, TIR: 90</td>
<td>02/07/00, 01/08/05, 26/06/06</td>
<td>Level 1B – 15 orthorectified (radiometric and geometric correction applied) images and DEM generated from forward (3N) and backward (3B) viewing channels</td>
<td>Vertical: +/- 25 m, Horizontal: +/- 10-30 m</td>
</tr>
<tr>
<td>Landsat 1 MSS</td>
<td>Orthorectified images – all 4 MSS channels</td>
<td>Visible (4-5): 80, NIR (6-7): 80</td>
<td>13/07/75</td>
<td>Orthorectified, with image to image registration</td>
<td>Positional accuracy: 100 m RMS</td>
</tr>
<tr>
<td>Landsat 5 TM</td>
<td>Orthorectified images – all 7 TM channels</td>
<td>Visible (1-3):30, NIR (4-5): 30, TIR (6): 120, Mid IR (7): 30</td>
<td>10/09/90</td>
<td>Orthorectified, with photogrammetric block adjustment</td>
<td>Positional accuracy: 50 m RMS</td>
</tr>
<tr>
<td>Landsat MSS</td>
<td>Wavelength (µm)</td>
<td>Spatial resolution (m)</td>
<td>Landsat TM and ETM+</td>
<td>Wavelength (µm)</td>
<td>Spatial resolution (m)</td>
</tr>
<tr>
<td>-------------</td>
<td>-----------------</td>
<td>------------------------</td>
<td>---------------------</td>
<td>-----------------</td>
<td>------------------------</td>
</tr>
<tr>
<td>Band 1 Visible</td>
<td>0.52-0.60</td>
<td>60x80</td>
<td>Band 1 Visible</td>
<td>0.45-0.52</td>
<td>30/30</td>
</tr>
<tr>
<td>Band 2 Visible</td>
<td>0.60–0.70</td>
<td>60x80</td>
<td>Band 2 Visible</td>
<td>0.52-0.60</td>
<td>30/30</td>
</tr>
<tr>
<td>Band 3 NIR</td>
<td>0.70-0.80</td>
<td>60x80</td>
<td>Band 3 Visible</td>
<td>0.63-0.69</td>
<td>30/30</td>
</tr>
<tr>
<td>Band 4 NIR</td>
<td>0.80-1.10</td>
<td>60x80</td>
<td>Band 4 NIR</td>
<td>0.77-0.90</td>
<td>30/30</td>
</tr>
<tr>
<td>Band 5 NIR</td>
<td>1.55-1.75</td>
<td>30/30</td>
<td>Band 6 SWIR</td>
<td>2.08-2.35</td>
<td>30/30</td>
</tr>
<tr>
<td>Band 7 Mid IR</td>
<td></td>
<td></td>
<td>Band 5 SWIR</td>
<td>2.145-2.185</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Band 6 SWIR</td>
<td>2.185-2.225</td>
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</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Band 8 SWIR</td>
<td>2.295-2.365</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Band 9 SWIR</td>
<td>2.360-2.430</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Band 10 TIR</td>
<td>8.125-8.475</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Band 11 TIR</td>
<td>8.475-8.825</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Band 13 TIR</td>
<td>10.25-10.95</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Band 14 TIR</td>
<td>10.95-11.65</td>
<td></td>
</tr>
</tbody>
</table>

**Table 4.6: Wavebands of the Landsat MSS TM and ETM+ and TERRA ASTER**
The correction of any deficiencies is known as ‘pre-processing’ and is carried out before any quantitative analysis is undertaken (Mather, 2004), with correction completed manually or by the supplier before the satellite imagery is obtained. However, an understanding of the processes completed to correct these data had to be considered, as different procedures can have very different impacts upon the raw data image values through the introduction of errors, bias, and loss of information. The importance of this was also highlighted in a previous investigation into the impact of different atmospheric correction methods applied to an ASTER image (Foster, 2006).

These images were acquired already radiometrically, atmospherically, and geometrically corrected (Table 4.5). Further details on how the Land Process Distributed Active Archive Centre (LP DAAC, who process the ASTER images) pre-processed these images can be found in Lang and Welch (1999), Arai and Tonooka (2005), Sakuma et al., (2005) and ASTER user guides available from Aster-web NASA (2008) and the Land Process Distributed Active Archive Centre (LP DAAC) (2008). Details on the pre-processing steps applied to the Landsat data by the Earthsat Corporation can be obtained from Landsat (2008).

4.3.1. ASTER DEM

ASTER DEM products were investigated to assess their potential for monitoring glacier surface changes. ASTER DEMs were utilised due as the application of other sensors such as radar, laser altimetry and SAR interferometry is hampered due to the problems of snow melt and influence of the extreme topography in mountainous regions (Kaab, 2007).

The ASTER sensor acquires simultaneous stereo images from different directions (forward and backward view), therefore, it can be used to generate a DEM with a 30 m resolution (Figure 4.13). To create an ASTER DEM, level 1A data are processed using the two stereo
bands of 3n (nadir viewing) and 3b (backward viewing) bands. This results in two views of
the same region from two vantage points and can be used to generate a DEM.

**Figure 4.13:** Generation of DEM through stereo image collection (ERSDAC, 2005).

Prior to 2006, both absolute (using GPS points of the study area in question) and relative
DEM could be ordered on demand from the EOS (Earth Observing System) data gateway.
However, since 2006, only relative DEMs are produced (which do not use any GPS points),
and are based upon an automated stereo correlation method. These relative DEMs are not tied
to a ground or map datum but to the lowest elevation in the scene, in comparison to an
absolute DEM which would be referenced to sea level (Hirano *et al.*, 2003). Therefore,
relative DEM generation utilises the ephemeris and altitude data recorded by both the ASTER
sensor, and TERRA satellite. This single band product is geodetically referenced to the UTM
co-ordinate system, and referenced to the Earth’s co-ordinate system through the EGM96
geopotential model. The accuracy of these later products are reported as being around 25 m
(Lang and Welch, 1999). Full details of DEM generation are found in the ASTER Users Guide (ERSDAC, 2005).

In this project four, DEMs acquired on 02/07/00, 14/08/04, 01/08/05, and 26/06/06 were used. These dates were selected after searching through all of the available images for the study area since the start of the ASTER mission in 2000. The most suitable images were those with no cloud/snow cover and, to avoid snow cover, obtained in July/August. However, the number of suitable images available for the study was limited, as many were completely obscured by cloud during the most suitable time of year, or were completely snow covered earlier/later in the year. Some of the images, however, do contain some cloud or snow patches which may impact upon the resulting DEM generation, but with no other alternative data available these were selected and the impacts of these lesser quality images was investigated through error analysis.

4.3.2. Orthophoto DEM

Two orthophoto DEMs were also available for use in analysis and were obtained in 1991 and 2003 by the Milan Earth Sciences Group at the University of Milan (Figure 4.14). These were generated using two stereo-pairs of aerial photographs, and had a spatial resolution of 10 m. Further details on the generation of the DEM can be found in D’Agata et al., (2005), and Diolaiuti et al., (2009). The 1991 DEM (vertical accuracy of ~0.36 m) covered all of the Miage Glacier and its tributaries. However, the 2003 DEM did not include the upper reaches and tributaries and, therefore, could not be used in this study. The 1991 DEM was re-sampled to a 90 m resolution (same spatial resolution as ASTER TIR) and overlain with the ASTER thermal image from 2005 so that the elevation and surface temperature could be obtained for every pixel on the image.
Figure 4.14: 1991 orthophoto DEM of the Miage Glacier and surrounding landscape.

4.3.2.1. Slope and Aspect

Information regarding the slope and aspect of every pixel was obtained from the DEM file using standard analysis methods and this was extracted into two separate layers (Figure 4.15 a and b). These were then added in as extra channels on the ASTER image, so that along with elevation, and surface temperature, the slope and aspect of each pixel was contained in one image. Once the ASTER image, DEM, and slope and aspect images (all at 90 m) had been combined into one image file, relevant pixels (the glacier area) were extracted in ASCII format. This meant that data values for the ASTER surface temperature, elevation, slope, and aspect for every pixel in each image at a 90 m resolution could be analysed in a spreadsheet format.
Figure 4.15: a) slope angle, b) slope aspect for each pixel in the DEM

4.4. Summary

This chapter has described both the study site and the equipment and methods used to collect the fieldwork data used throughout the remainder of this thesis. Field data included surface
temperature and air temperature measurements, debris thickness measurements, rock lithology observations, and monitoring of possible causes for large surface elevation changes.

It has also provided a background to the historical evolution of the Miage Glacier over the Holocene. Knowing and understanding the causes of historical changes on the glacier means the findings of later chapters in this study have a basis for comparison, and whether results or patterns that are identified have occurred previously.

Finally, an introduction to the satellite images used is included as well as the identification of any pre-processing completed upon them. Understanding the processes that have been applied to the satellite images means that analysis can be completed successfully and implications of the methods applied, such as atmospheric correction or calculation of surface temperatures, considered in each of the applications.
CHAPTER 5: EXTRACTING DEBRIS THICKNESS FROM THERMAL BAND IMAGERY

Chapter aim: To develop a method to estimate debris thickness from ASTER thermal-band imagery using a physically-based energy balance model.

5.1. Introduction

Debris thickness is one of the key variables required to calculate ablation over a glacier-wide area (Mihalcea et al., 2008). Field measurements are time-consuming and difficult in many locations, highlighting the potential applicability of remote sensing to resolve the problem. However, previous attempts to estimate debris thickness from satellite imagery have been limited, despite the potential advantages of using such a method, including rapid data acquisition, ability to access difficult to reach locations on the ground, and the possibility to estimate debris thicknesses at higher spatial and temporal frequencies than ground measurements. Therefore, a method that estimates debris thickness on a glacier using a thermal surface temperature image was investigated and developed, with ASTER thermal band imagery being used to overcome limitations of using visible and infra red satellite imagery in estimating debris cover on glaciers (Taschner and Ranzi, 2002).

This chapter, therefore, focuses upon the generation and testing of this model, and is split into sections looking at the issues of both the validation of ASTER surface temperature data and the relationship between debris thickness and surface temperature using field measurements, along with the extraction of debris thickness estimates using both empirical and energy balance approaches. The first section is focused on the validation of ASTER satellite surface temperatures through ground based thermistor measurements, along with the identification of a
relationship between surface temperature and debris thickness. Factors affecting the spatial
variability of surface temperature independently of thickness were also investigated including
shading, $SWR_\downarrow$ and variable lithology. The variability of debris thickness on the glacier
surface was investigated through spatial autocorrelation techniques to determine the ‘ideal’
spatial resolution for measurement and whether it corresponded to the 90 x 90 m resolution of
the ASTER image.

Subsequent sections are focused on the development of a method to determine debris thickness
from thermal satellite imagery. Previously, studies such as Mihalcea et al. (2008) have utilised
field data (measured debris thicknesses and surface temperatures) to develop empirical
approaches based upon the relationship of debris thickness to surface temperature. By
developing an empirical equation, debris thicknesses can be estimated for an entire debris
covered part of a glacier, by substituting ASTER thermal imagery surface temperatures into
the equation. These empirical approaches are investigated and their applicability and
transferability tested on the Miage Glacier.

A second approach is developed here, focused on the development of a physically-based
energy balance approach to estimate debris thickness, utilising surface temperature from an
ASTER image and measured meteorological values. The final part of this chapter is focused
on the comparison of the different methods attempted and an assessment of their accuracy
Therefore, remotely sensed data is utilised to obtain surface temperature data, and as a result
estimate debris thickness, and field data is utilised in the energy balance model (AWS data),
and to validate the model debris thickness estimates.
5.2. Methodology

A number of different processes were applied during the development of a method to estimate debris thickness on a debris-covered glacier, these are clearly outlined below.

1) Validation of ASTER surface temperature with field based measurements, including analysis of spatial variability of surface temperature in the field over an area the size of an ASTER pixel.

2) Identification of a surface temperature – debris thickness relationship.

3) Investigation of variability of debris thickness at different locations on a glaciers surface.

4) Application of an empirical method to obtain debris thickness from ASTER imagery, based on the relationship between debris thickness and surface temperature.

5) Application of an energy balance model to calculate surface energy fluxes, to ensure it could be successfully applied at the Miage Glacier.

6) Development of the energy balance model to enable calculation of debris thickness from surface temperature obtained from an ASTER image. To enable debris thickness estimations at all pixels on the glacier an equation to estimate air temperature was developed based on the air temperature – surface temperature relationship.

7) Debris thickness map generated.

8) Comparison of results to field data and a previous study on the Miage Glacier using an Empirical approach (Mihalcea et al., 2008).

9) Sensitivity analysis to input parameters and development of the energy balance model, making it more physically-based by calculating $SWR_{\downarrow}$ and $LWR_{\uparrow}$ for each pixel using a DEM, to enable more accurate determinations of debris thickness.

10) Application of the energy balance model to an ASTER image from another year (2004).

5.3. Validation of ASTER with field observations

The method of debris thickness estimation is based on the principle that the thicker the debris, the higher the surface temperature. Therefore, if surface temperature can be obtained for the entire debris-covered area, debris thickness can be determined, highlighting the potential of remote sensing for providing the surface temperature values. However, before this method can be developed two aims need to be addressed. Firstly, the validation of satellite surface temperatures, and secondly, the identification of the relationship of surface temperature to debris thickness. Surface thermistor data and debris thicknesses collected in the field were used to achieve these aims. The applicability of these thermistor measurements is also assessed through the investigation of how representative individual thermistor measurements are of the 90 x 90 m area within an ASTER pixel. This was achieved through the investigation into the impact of different microenvironments, due to variable microtopography and lithology, along with weather conditions, on thermistor temperature measurements.

5.3.1. Surface thermistor temperatures vs. ASTER temperatures

To validate the satellite derived surface temperatures from the ASTER image, values were compared against surface temperatures derived from 21 surface thermistors installed on the debris surface in 2005 at the time of image acquisition (01/08/05, 10:40). These thermistors were located across the glacier surface in a range of debris thicknesses, from 0.04 m to 0.5 m, and at a range of different elevations (1839 m to 2419 m).

Figure 5.1 shows a relationship is present between the two variables, which is supported by a correlation of $r = 0.695$. However, it is clear that the thermistors tend to over estimate surface
temperature compared to ASTER, particularly in cooler (thinner) areas of debris. A reason for this scatter could be due to the positioning of the thermistor probes on the top of flat up-facing clasts, meaning they are not necessarily representative of the surrounding debris (especially in areas of thin/sparse debris cover) (Mihalcea et al., 2008). Whereas the ASTER surface temperatures provide an average temperature value over a 90 x 90 m area, highlighting the issue of spatial scale in surface temperature measurements.

![Figure 5.1: ASTER surface temperatures vs. thermistor surface temperatures (01/08/05 10:40) at 21 sample points and corresponding ASTER pixels.]

5.3.2. Spatial variability of surface temperatures

To test the spatial variability of surface temperatures over an area the size of an ASTER pixel (90 x 90 m), and to see how representative surface thermistors are in relation to the ASTER pixels, a number of thermistor transects were installed on the glacier surface (details of thermistor set up and transect locations can be found in section 4.4.2). By setting out these transects, the temperature variability over 2 m, 5 m, and 20 m spacings in both across and down glacier directions could be established. The dependence of surface temperature on microtopography and rock type in an area of constant debris thickness was also assessed. To
identify the amount of spatial variability surface temperatures from each of the thermistors for a single day at 10:40 (time of ASTER image acquisition), and the average temperatures for the sampling period were plotted against distance along the transect. The 21/06/07 was selected for the 2 m spacing, 23/06/07 for the 5 m, and 31/08/07 for the 20 m. All days selected had high $SWR\downarrow$ to ensure similar conditions to the day of image acquisition, and were at least a day after the surface thermistors were set up to allow the debris temperature profile time to adjust following the disturbance of the debris during thickness measurement.

The data from the 2 m, 5 m, and 20 m transects (Figure 5.2 a-c, and Table 5.1 a and b) show high spatial variability in surface temperatures for a single day at 10:40. Although variability is much more apparent at 5 m and 20 m spacings, with temperature variability just $\sim3$ °C for the 2 m spacing, but $\sim19$ °C for the 5 m, and $\sim10$ °C for the 20 m. The variability on the smaller transects (2 m and 5 m) is largely the result of the surface microenvironment and its interaction with incoming solar radiation, due to aspect, shading (e.g. low temperatures of 4.5 °C and 6.7 °C), and rock albedo. The differences in temperature over the larger 20 m scale can be further explained by the presence of a number of factors including ice cliffs, variable debris depths, different rock types, and exposed ice locations (Mihalcea et al., 2008).

The temperatures experienced at 10:40 on the 21/06/07, 23/06/07, and 31/08/07 are higher than the average surface temperatures between 20/06/07-21/06/07, 22/06/07-17/07/07, and 31/08/07-04/09/07. Lower average surface temperatures at all spatial scales are experienced due to the inclusion of data from both days and nights (rather than a single measurement at a specific time during the day), during which different levels of $SWR\downarrow$ (with its impact upon surface temperatures) will have been received. Therefore, spatial variability in temperature at 2 m, 5 m, and 20 m scales is not as prominent in longer-term average temperature data (Table 5.1 and Figure 5.2) and is almost absent at night (Figure 5.2 d).
Figure 5.2: Surface temperature values in comparison to daily 10:40 temperatures and average values for the sample period a) 2 m spacing glacier transect (20/06/07 - 21/06/07), b) 5 m spacing glacier transect (22/06/07 – 17/07/07), c) 20 m spacing glacier transect (31/08/07 – 04/09/07), d) Average surface temperature during the night (00:00-06:00) – of the 5 m across and down glacier transect.
Table 5.1: Summary of surface temperatures for all spatial scales, a) 2 m and 5 m small scale transects, b) large scale 20 m transects

a)

<table>
<thead>
<tr>
<th>Transect spacing (m)</th>
<th>Transect length (m)</th>
<th>$T_s$ average (°C)</th>
<th>Standard deviation</th>
<th>Min $T_s$ (°C)</th>
<th>Max $T_s$ (°C)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Average 20/06/07-21/06/07</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Down glacier</td>
<td>2</td>
<td>12</td>
<td>17.1</td>
<td>2.0</td>
<td>13.9</td>
</tr>
<tr>
<td>Across glacier</td>
<td>2</td>
<td>6</td>
<td>16.4</td>
<td>0.5</td>
<td>16.1</td>
</tr>
<tr>
<td>21/06/2007 10:40</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Down glacier</td>
<td>2</td>
<td>8</td>
<td>18.3</td>
<td>1.1</td>
<td>16.7</td>
</tr>
<tr>
<td>Across glacier</td>
<td>2</td>
<td>4</td>
<td>16.2</td>
<td>0.2</td>
<td>16.0</td>
</tr>
<tr>
<td>Average 22/06/07-17/07/07</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Down glacier</td>
<td>5</td>
<td>30</td>
<td>11.4</td>
<td>0.5</td>
<td>10.4</td>
</tr>
<tr>
<td>Across glacier</td>
<td>5</td>
<td>25</td>
<td>11.7</td>
<td>0.3</td>
<td>11.3</td>
</tr>
<tr>
<td>23/06/2007 10:40</td>
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<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Down glacier</td>
<td>5</td>
<td>30</td>
<td>17.4</td>
<td>7.3</td>
<td>4.5</td>
</tr>
<tr>
<td>Across glacier</td>
<td>5</td>
<td>20</td>
<td>16.2</td>
<td>6.4</td>
<td>6.7</td>
</tr>
</tbody>
</table>

b)

<table>
<thead>
<tr>
<th>Transect spacing (m)</th>
<th>Transect length (m)</th>
<th>$T_s$ average (°C)</th>
<th>Standard deviation</th>
<th>Min $T_s$ (°C)</th>
<th>Max $T_s$ (°C)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Average 31/08/07-04/09/07</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Across glacier</td>
<td>10</td>
<td>100</td>
<td>9.1</td>
<td>1.7</td>
<td>7.0</td>
</tr>
<tr>
<td>Down glacier</td>
<td>10</td>
<td>100</td>
<td>11.1</td>
<td>0.9</td>
<td>9.7</td>
</tr>
<tr>
<td>31/08/2007 10:40</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Across glacier</td>
<td>10</td>
<td>100</td>
<td>21.2</td>
<td>3.9</td>
<td>16.7</td>
</tr>
<tr>
<td>Down glacier</td>
<td>10</td>
<td>100</td>
<td>26.2</td>
<td>3.4</td>
<td>22.0</td>
</tr>
</tbody>
</table>

These findings support those of Mihalcea et al., (2008), on the Miage Glacier, who highlight that at a scale <1 day surface temperature is influenced by the surface microtopographical conditions such as shading, roughness, elevation, lithology, and their interaction with $SWR\downarrow$, whereas over a period greater than this, the debris thickness becomes the primary control over surface temperature. The influence of these factors is investigated further in section 5.3.3 and ultimately results in problems relating to the validating of satellite temperatures with in-situ thermistor measurements, especially when the number of surface thermistors and their locations are not representative of an ASTER pixel area (90 x 90 m). These limitations with
surface thermistors due to small scale variability in surface temperatures highlight that satellites are a better way to measure surface temperature at the glacier scale, and that satellite derived surface temperatures are better for use in debris thickness estimations over large spatial scales, since any variations in surface temperature associated with different microtopographies will be averaged out within a pixel. However, thermistors are the best method for surface temperature retrieval at a specific location on a glacier's surface.

Overall, analysis of the spatial variability of surface temperature has identified that the direct measurement of surface temperature is very sensitive to the actual location of the thermistors, due to individual thermistor measurements being affected by the surrounding microtopography, variable lithology, and the influence of the sun and wind which can either heat or cool the sensors. To address this issue, field based thermistor studies on surface temperature at a large spatial scale must sample a range of microtopographies in different locations (including shaded hollows, different lithologies, and different slope angles/aspects) over a large area, which would be difficult to complete over a large glacier or large number of glaciers. This highlights the advantage of satellite derived surface temperatures for large spatial scales, as the impact of small scale spatial variability of surface temperatures is averaged out within an ASTER satellite pixel’s 8100 m² area.

5.3.3. Analysis of the effects of different microenvironments on thermistor temperature

Generally, surface temperature graphs follow a characteristic diurnal pattern on a cloudless summer day, where in the morning most of the energy is arriving much faster than it can be dissipated (Figure 5.3). This means that the input is exceeding the output and results in an accumulation of energy, which subsequently causes the temperature of the surface to increase (Oke, 1987).
Figure 5.3: Relationship between the surface energy exchange and the diurnal cycle of surface temperature (Oke, 1987).

![Graph showing energy flux density and surface temperature over time.]

Figure 5.4: Characteristic diurnal pattern of surface temperature – average temperatures calculated from the entire 2005 data season (06/06/05 – 08/09/05), recorded at the LWS.

The time that this occurs on the Miage Glacier is shown by the rapid rise in temperatures at 07:30, where the presence of the Sun for a number of hours has resulted in the amount of incoming energy exceeding the output energy (Figure 5.4). The top of the steep rise on the graphs shows the maximum temperature, indicating when both the input and output energy quantities are equal (occurring at 14:00, which is effectively solar noon at the location of the Miage Glacier in the summer). From this point on, more heat is extracted than is being added,
resulting in a drop in temperature (Oke, 1987). Fluctuations may occur throughout the day, when the output briefly exceeds the input, and can occur if clouds are present, which reduce the amount of $SWR_\downarrow$ reaching the ground, and results in surface cooling while the cloud is present then heating once it has passed (Figure 5.5).

![Figure 5.5: Diurnal $T_s$ and $SWR_\downarrow$ 01/08/05 showing fluctuations between 07:30-20:00 at LWS site.](image)

The temperature starts to drop on the Miage around 16:00. Here the rate of loss is greater than the rate of gain, and is experienced towards the end of the day because the Sun’s energy is reduced as it sets. This pattern is repeated each day as the Sun rises and sets, resulting in a diurnal surface temperature regime which is experienced across the glacier surface. The same cycle is also repeated beneath the surface, with the range in surface temperatures decreasing with an increase in depth (Figure 5.6).
Variations in this diurnal cycle occur due as a result of: i) locations which are shaded from the sun at certain points during the day (Figure 5.9), ii) different rock types which heat up differentially (Figure 5.8), iii) different debris thicknesses (Figure 5.10). Local aspect is also important, due to differential warming, where west facing slopes will experience highest surface temperatures in late afternoon, compared to easterly facing slopes where they will be highest during the mid morning. Therefore, the impacts of these different factors will be investigated below.

5.3.3.1.Rock underside

When one thermistor is placed on the underside of a rock it is clear that temperatures on top of the rock are warmer than those experienced on the bottom (Figure 5.7, Table 5.2).
Figure 5.7: a) location of thermistors b) average diurnal temperature variability between 31/08/07-04/09/07 of a thermistor sensor placed on the top of a rock and another placed on the underside, clast thickness ~0.02 m (thermistor located 100 m from the LWS).

This pattern was expected due to the shading of the base from the Sun’s rays. However, between 13:00 and 15:00 a peak in base thermistor temperatures occurs, whilst the top of rock thermistors temperatures are declining. This drop may be due to the common occurrence of cloud cover over the Miage Glacier during the afternoon, which reduces the $SWR\downarrow$. Therefore, the surface rapidly loses heat through radiation and sensible heat. In contrast, the underside sensor remains warm due to heat conducting from the surface throughout the day, highlighting
that temperature at depth is more stable, and increases as the heat is being conducted down
from the surface during this time.

Table 5.2: Example temperature values from the upper and underside of a rock at 10:40 (time of ASTER image acquisition)

<table>
<thead>
<tr>
<th>Date/Time</th>
<th>Upperside (°C)</th>
<th>Underside (°C)</th>
<th>Temperature difference (°C)</th>
</tr>
</thead>
<tbody>
<tr>
<td>01/08/07 10:40</td>
<td>17.40</td>
<td>12.40</td>
<td>-5.00</td>
</tr>
<tr>
<td>02/08/07 10:40</td>
<td>19.50</td>
<td>14.50</td>
<td>-5.00</td>
</tr>
<tr>
<td>03/08/07 10:40</td>
<td>19.10</td>
<td>13.50</td>
<td>-5.60</td>
</tr>
<tr>
<td>04/08/07 10:40</td>
<td>9.50</td>
<td>8.80</td>
<td>-0.70</td>
</tr>
<tr>
<td>05/08/07 10:40</td>
<td>13.80</td>
<td>9.50</td>
<td>-4.30</td>
</tr>
</tbody>
</table>

5.3.3.2. Comparison of rock type

When looking at the impact of different rock types a clear pattern of the darker schist (lower albedo) being slightly warmer, and granite (higher albedo) being slightly cooler is evident (Figure 5.8, Table 5.3). This is due to the lower albedo of the darker schist which, as a result, absorbs more solar radiation, than the higher albedo of the granite, which reflects more solar radiation. Therefore, different debris thicknesses could be inferred (from satellite imagery) at a site where two different rock types are present, even if the actual depths are the same.

Table 5.3: Example temperature values from rocks with different albedo (one a dark schist, the other a lighter granite) at 10:40 (time of ASTER image acquisition)

<table>
<thead>
<tr>
<th>Date/Time</th>
<th>Upperside (°C)</th>
<th>Underside (°C)</th>
<th>Temperature difference (°C)</th>
</tr>
</thead>
<tbody>
<tr>
<td>01/09/07 10:40</td>
<td>25.9</td>
<td>19.1</td>
<td>-6.8</td>
</tr>
<tr>
<td>02/09/07 10:40</td>
<td>30.3</td>
<td>23.7</td>
<td>-6.6</td>
</tr>
<tr>
<td>03/09/07 10:40</td>
<td>27.4</td>
<td>21.3</td>
<td>-6.1</td>
</tr>
<tr>
<td>04/09/07 10:40</td>
<td>15.3</td>
<td>12</td>
<td>-3.3</td>
</tr>
</tbody>
</table>
5.3.3.3. Shading

When looking at the impact of the thermistor sensor placed near an area of shadowing (caused by the surrounding rocks), it is clear that the temperature of the rocks is reduced as it is shaded away from direct sunlight which would heat the rock’s surface (Figure 5.10). Shadow can have an impact upon the debris thickness calculated at a location when using surface thermistors, as a thick debris thickness would be under-estimated due to the lower surface temperature recorded by the thermistor.

Figure 5.8: a) Tiny_Tag 13_14 at the AWS site, b) ‘typical’ diurnal temperature variations (31/08/07-04/09/07) between thermistors on different rocks, one a dark schist, the other a lighter granite.
Figure 5.9: Diurnal temperature variations between a thermistor paced in a shaded location and the other in an open location, a) thermistor location, b) average hourly surface temperatures (31/08/07-04/09/07).

However, if the sample area is large enough – as in the case of an ASTER pixel (90 x 90 m) the debris thickness estimation is unlikely to be affected by shadow, due to the averaging out of these shaded areas within a pixel. The impact of shading on recorded surface temperatures highlights the advantage of using satellite measurements of surface temperature over a large spatial scale. Also the presence of shadow provides an explanation for why thermistor surface temperatures (point measurements) are higher than ASTER (average 90 x 90 m areas) surface temperatures for the same location (Figure 5.1). Because the surface thermistors were not
located in an area of shadow, whereas, the ASTER pixels will have included some areas of shade and will therefore record a lower surface temperature.

### 5.3.3.4. Varied depths

When placed on two different debris thicknesses (0.05 m, 0.4 m) it is clear to see that thicker debris layers have much warmer temperatures (Figure 5.9), and this is what enables the identification of debris thickness.

![Figure 5.10: Average daily temperature variations between thermistors on debris thicknesses 0.4 m and 0.05 m a) location of thermistors, b) average hourly surface temperatures for the sampling period (31/08/07-04/09/07).](image-url)
The debris layer of 0.05 m rarely exceeds temperature of 15 °C, and experiences peak temperatures between 10:00-12:00, when SWR↓ is at its highest. The thicker debris is much warmer (and is above 15°C between 10:00 and 19:00) due to its greater thermal resistance, and there being more material to heat up and store heat compared to thinner thicknesses.

Overall, when the impact of sensor location and albedo are compared both have a similar effect upon surface temperatures recorded, with the range of temperature differences identified between rocks with different albedo values being of a similar magnitude to those differences obtained when sensors are placed in different locations such as shadow.

### 5.4. Surface temperature-debris thickness relationship

#### 5.4.1. Surface data from thermistors

The relationship between surface temperature and debris thickness was investigated using data collected in 2005 and 2007 (surface temperatures obtained from thermistors), and the relationship between debris thickness and surface temperature at 10:40 (the time of ASTER image acquisition) compared to see if and how the relationship varies. Any changes that occurred in the relationship at different times of the day and night including 06:00-12:00 and 00:00-03:00 were also investigated to identify the time at which the relationship is at its strongest.

<table>
<thead>
<tr>
<th>Debris thickness vs. thermistor temperature</th>
<th>Pearson’s correlation (r)</th>
<th>P-value</th>
</tr>
</thead>
<tbody>
<tr>
<td>06:00-12:00 01/08/05</td>
<td>0.533</td>
<td>0.013</td>
</tr>
<tr>
<td>00:00-03:00 01/08/05</td>
<td>0.775</td>
<td>0.000</td>
</tr>
<tr>
<td>10:40 2005</td>
<td>0.254</td>
<td>0.267</td>
</tr>
<tr>
<td>2007</td>
<td>0.369</td>
<td>0.264</td>
</tr>
</tbody>
</table>
Figure 5.11: Surface temperature (from surface thermistors) vs. debris thickness at a) 10:40 01/08/05 (time of ASTER image acquisition) in 2005 (stake locations – over entire glacier) and b) 10:40 31/08/07 (debris transect locations middle glacier), c) average surface temperature between 06:00-12:00 01/08/05, d) average surface temperature between 00:00-03:00 01/08/05.
The strength of the temperature – thickness relationship increases between 06:00-12:00 (r = 0.533) in 2005 and is at its greatest between 00:00-03:00 (r = 0.775) in 2005 (Figure 5.11 c and d, Table 5.2). A reason for the increase in the strength of relationship between surface temperature and debris thickness is due to it being at night, meaning it is unaffected by $SWR_↓$. $SWR_↓$ has an impact, during the daytime, upon the strength of the temperature – thickness relationship, as it differentially warms the debris layer, due to the presence of different rock types and debris thicknesses which warm at different rates. However, despite an increase in the temperature-thickness relationship, some anomalies are still present on the 00:00-03:00 plot (Figure 5.11 d), and may occur due to the differential cooling of the debris during the night, because, some rocks types will stay warmer than others for a longer duration due to their thermal properties.

Anomalies are also present on the other plots (Figure 5.11a, b, and c) and may occur due to the use of a point debris thickness measurement at the thermistor locations which may not be accurate or representative of the surrounding debris area. Also, elevation will have an impact upon the temperature-thickness relationship as the air is warmer lower down the glacier, meaning the relationship between surface temperature and debris thickness will vary at different elevations. The positioning of the surface thermistors on the glacier surface will also have an impact, if any sensors become shaded for example, the relationship between surface temperature and debris thickness will be affected, as cooler surface temperatures will be recorded for that debris thickness compared to an unshaded sensor at a similar debris thickness.

5.4.2. Debris thickness variability

To identify the extent of the spatial autocorrelation of debris thicknesses at different locations on the glacier, semivariograms were completed using measured debris thicknesses from across
and down glacier transects collected in 2006 and 2007 (details on the location of these transects is found in Figure 4.14). Using the level of spatial autocorrelation highlighted by the semivariograms, the variability in debris cover thickness in relation to the spatial resolution of the image data could be identified.

Results show (Table 5.3) that the sill values (the maximum distance at which spatially autocorrelated measurements can occur) obtained for the lower, middle, and upper transects in both across and down glacier directions are all relatively low, with the highest sill being only 30 m. In turn, this indicates that there is a low spatial autocorrelation of debris thicknesses throughout the glacier.

<table>
<thead>
<tr>
<th>Location</th>
<th>Across glacier</th>
<th>Down glacier</th>
</tr>
</thead>
<tbody>
<tr>
<td>Lower glacier 2006</td>
<td>14</td>
<td>20</td>
</tr>
<tr>
<td>Middle glacier 2006</td>
<td>21</td>
<td>12</td>
</tr>
<tr>
<td>Upper glacier 2006</td>
<td>NA</td>
<td>12</td>
</tr>
<tr>
<td>Upper glacier 2007</td>
<td>30</td>
<td>10</td>
</tr>
</tbody>
</table>

Low spatial correlation of debris thickness occurred in the upper reaches due to the presence of bare ice patches (Figure 5.12 b), ice cliffs, sparse debris covers (Figure 5.12 b), and medial moraines which consisted of thicker debris. All of which combine to result in highly varied debris thicknesses along a transect (Figure 5.12 e.).
Figure 5.12: Photographs of the upper transect a) 2006, b) 2007 arm 1, c) 2007 arm 3, showing transect was taken along the top of steep sided moraine, d) 2007 arm 4, taken up towards the top of a steep ice cliff, e) 2007 arm 2, transect from the top of the steep moraine into the trough area. Red lines show transect direction.
Figure 5.13: Photos of the middle transect a) arm 1, across glacier, showing transect was taken over a moraine, b) arm 2, across glacier, also across a moraine, c) arm 3, down glacier d) arm 4, up glacier, along a steep moraine also showing large number of large boulders on the top of the moraine. Red lines show transect direction.

A reason for these low values in the lower and middle reaches are the presence of moraines on the glacier surface (mentioned previously in section 2.2.1), with thicker debris found at the top of these moraines, and thinner debris found at the base. Therefore, variable debris thicknesses result when a transect is taken across one of these moraines, such as the lower across glacier transect (Figure 5.14 a), compared to the down glacier transects which were taken along the top of one of these moraines (Figure 5.14 c). The presence of large boulders on the debris surface also generates large thickness variability over a small area, with thicknesses ranging from a few centimetres to over half a metre over very short distances (Figure 5.13 b and d).
Figure 5.14: Photographs of the lower transect, a) arm 1, across glacier, showing samples were taken across a moraine, b) arm 2, c) arm 3, up glacier, showing transect was taken along the top of a moraine, d) arm 4. Transect direction shown by red lines.

Overall, therefore, the suggested spatial resolution for analysis of this debris thickness varies between 14-21 m in the lower and middles reaches, to 30 m in the upper reaches, confirming an assumption that the 90 x 90 m resolution ASTER image is too coarse to capture all of the spatial variability in debris thickness. However, since the ASTER temperature for the 90 x 90 m pixel area will be an average for the entire pixel area it will include all of the thickness variations and, therefore, produce an average thickness value for that 90 x 90 m area.
5.5. Estimating debris thickness with ASTER imagery

5.5.1. Empirical method

In an attempt to estimate debris thickness from thermal satellite imagery, first, an empirical approach was used whereby the relationship between debris thickness and surface temperature was developed using regression analysis. From this, the estimation of debris thickness through regression equations/analysis could be identified when only surface temperature was known (from the ASTER image). Empirical approaches can only be utilised when extensive field data is available to develop the relationship, and have previously been applied successfully to other debris-covered glaciers such as the Baltoro Glacier, Karakoram, Pakistan (Mihalcea et al., 2008), and the Miage Glacier (Mihalcea et al., 2008).

5.5.1.1. Debris temperature relationship

To develop an empirical approach on the Miage Glacier, data from 01/08/05 at 10:40 was selected due to the ASTER image being acquired at this date and time. Surface temperatures acquired by the surface thermistors and the surface temperatures from the ASTER pixels corresponding to the thermistor locations were utilised in this analysis.

Firstly, linear relationships were fitted (Figure 5.15a and b), and results show that similarly to section 5.3.1, the data are not tightly clustered around the line of fit, highlighting that a strong linear relationship is not evident. Therefore, due to the independence of surface temperature from debris thickness beyond ~0.4 m (Ranzi, et. al. 2004) quadratic relationships were fitted (Figure 5.15 c and d) showing a much better fit to the data. However, although a better fit is obtained, the relationship itself is unrealistic over the full range of temperatures, because, a decrease in debris thickness is predicted for an increase in surface temperature between 12 and 20 °C.
**Figure 5.15:** Thermistor temperatures plotted against debris thickness, a) linear line of fit applied 01/08/05 at 10:40 (thermistor temperatures), b) linear line of fit applied 01/08/05 at 10:40 (ASTER temperatures), c) quadratic line of fit applied 01/08/05 at 10:40 (thermistor temperatures), d) quadratic line of fit applied 01/08/05 at 10:40 (ASTER temperatures).
This problem could be removed based on the assumption that surface temperature is 0 °C at 0 m in thickness, by using a linear fit between a depth of 0 m at 0 °C and a depth of 0.08 at 20 °C. Also, because the ASTER temperatures were for the pixel corresponding to the field measurement of debris thickness at a point location, there is a scale issue, since the thickness values from the field are not representative of the ASTER pixels 90 x 90 m area. Therefore, the relationship (and strength of the relationship) between ASTER surface temperatures and the measured debris thicknesses was affected. Other lines of fit were also investigated, including exponential and logarithmic. Results showed that the exponential provided similar results to the quadratic approach although lines were straighter and almost linear on some graphs, and the logarithmic implied that, in some cases, debris thickness decreased with an increase in temperature which was clearly incorrect. Therefore, these relationships were not used.

Figure 5.16: Plots of estimated debris thicknesses against ASTER pixel temperature obtained for both linear and quadratic regression equations, with linear extension to the quadratic plot for shallow debris depths.
Both the linear (Equation 5.1) and quadratic (Equation 5.2) regression equations (of the ASTER 10:40 values vs. debris thickness) were applied to each pixel in the ASTER image (Figure 5.16) to generate maps of debris thickness. The results are assessed through comparison to debris thicknesses obtained in the field in 2005, 2006, and 2007 (section 5.8.2).

\[ d = -0.0773 + 0.0090 \times ASTER \; T_s \]  \hspace{1cm} (5.1)

\[ d = 0.2469 - 0.0222 \times ASTER \; T_s + 0.0007 \times ASTER \; T_s \; ^2 \]  \hspace{1cm} (5.2)

The empirical approach applied here consisted of one single equation for the entire glacier. However, another approach which has been applied on the Miage Glacier uses separately calibrated linear equations based upon different elevation bands (Mihalcea et al., 2008), to account for the influence of elevation upon the debris surface temperature relationship which results in a lower surface temperature at a higher elevation for a specific debris depth. Therefore, the use of separate equations addresses this, with 8 used in total by Mihalcea et al., (2008). This approach, however, is specific to the glacier, time, and date, which has an impact upon its transferability, because the specific conditions on the Miage Glacier which these separate equations address differ at other locations. Despite this, it does highlight the potential of ASTER imagery to generate information on debris thickness (Mihalcea et al., 2008).

To ensure transferability to other locations and due to the complicated relationship between debris thickness and surface temperature during the daytime as a consequence of spatially variable SWR↓, due to slope and aspect variations and the influence of elevation (decrease in air temperature with increase in elevation and its impact on the turbulent fluxes and longwave radiation), an energy balance approach is needed, which can account for all of the fluxes of energy which influence surface temperature.
5.6. Energy balance modelling

Energy balance modelling was applied due to the unsuccessful application of an empirical approach for estimating debris thickness. An energy balance approach also has the potential of greater transferability at different locations and times, compared to the regression approach which relies upon equation developed using significant amounts of field data collected specifically at one glacier and a certain time and day. However, prior to the development of the energy balance model to estimate debris thickness, an energy balance model had to be applied to the Miage Glacier to ensure it could be applied to successfully estimate the surface energy balance fluxes.

To determine the surface energy balance of the debris on the Miage Glacier, a physically-based method was applied that has been used and tested previously on other debris-covered glaciers (Brock et al., 2007) (Equation 2.9 section 2.8). The bulk aerodynamic profile method with Richardson number was used to calculate the $SHF$ to account for variations in atmospheric stability, and $LHF$ and $PRE$ were ignored as the surface was dry at the time of image acquisition. The bulk aerodynamic profile method with Richardson number was used due its advantage of being relatively simple to apply. It also covers a range of atmospheric conditions (stable and unstable) without the need to change parameter values (unlike the Monin-Obukhov method), and has previously been successfully applied to both debris-free (e.g. Favier et al., 2004) and debris-covered (e.g. Brock et al., 2007) glaciers. Details of this method, along with the details of the calculation of each of the fluxes can be found in chapter section 2.7.

5.6.1. Energy balance modelling: application

Before the energy balance approach was used to calculate debris thickness, an investigation into the surface energy fluxes on the 01/08/05 was completed to gain an understanding of their
patterns. To compute the energy balance model for the Miage Glacier at the LWS, a simplified version of Equation 2.9 was used (Equation 5.3).

\[ SWR↓ + LWR↑ + SHF + ∂STOR = COND \quad (5.3) \]

Fluxes in Equation 4.3 were evaluated for 01/08/05 (date of ASTER image) using data at 10 minute intervals from the LWS (further details on the collection of data at the LWS are found in section 4.4.1., and the calculation of fluxes are found in section 2.7). A \( z_o \) value of 0.001 m was used following Brock et al., (2007), although this value is low considering the rough debris surface. An average thermal conductivity (\( K \)) value for the glacier of 0.96 Wm\(^{-1}\) K\(^{-1}\) was calculated using Equation 2.1 from 25 point locations at each of the ablation stake locations between June and September in 2005 (Brock et al., in press). \( LHF \) was not calculated because, first, the debris was dry at the time of image acquisition, therefore evaporation, and condensation at the surface was assumed to be negligible. Second, as surface temperature was measured, any energy used or gained by evaporation or condensation within the debris had already been accounted for.

The only net input of energy into the debris is \( SWR↓ \), and the largest output flux is \( SHF \) (Figure 5.17). These peak at 933Wm\(^{-2}\) and 423Wm\(^{-2}\), respectively. All of the fluxes peak during the middle part of the day, due to the peak in \( SWR↓ \) and resulting debris warming at this time therefore, sub-debris melting will be greatest at this time (Figure 5.18).
Figure 5.17: Plotted results of different surface energy fluxes at 10 minute intervals, with ∂STOR calculated hourly 01/08/05 measured at the LWS.

Table 5.6: Mean flux values over a 24 hour period (01/08/05)

<table>
<thead>
<tr>
<th>Flux</th>
<th>24 hour average flux (Wm⁻²)</th>
</tr>
</thead>
<tbody>
<tr>
<td>SWR</td>
<td>212.10</td>
</tr>
<tr>
<td>LWR</td>
<td>-72.96</td>
</tr>
<tr>
<td>SHF</td>
<td>-65.73</td>
</tr>
<tr>
<td>COND</td>
<td>-88.07</td>
</tr>
<tr>
<td>∂STOR</td>
<td>0.49</td>
</tr>
<tr>
<td>Residual</td>
<td>-14.17</td>
</tr>
</tbody>
</table>

There is a steady increase in SWR↓ up to 800 W m⁻² around the time of ASTER image acquisition (11:00) after this SWR↓ is highly variable in response to cloud cover variations.

This highlights clear sky conditions at the time of ASTER image acquisition (10:40), indicating its suitability for use in debris thickness estimations on the 01/08/05. SHF gradually increases to a value around 300 W m⁻² at the time of image acquisition (10:40), after this time SHF is highly variable due to changes in windspeed throughout the day. The COND values
peak at 200 Wm\(^{-2}\) between 10:00 and 18:00, highlighting that at this time more heat is conducted through the debris into the ice beneath therefore, this corresponds to the period with the largest levels of melt identified in Figure 5.18.

![Melt rate graph](image)

**Figure 5.18:** Melt rate in water equivalent units as a result of COND at hourly timescales on the 01/08/05.

The greatest changes in heat storage within the debris layer occur between 09:00-18:00, with a positive increase in the amount of stored heat occurring during the morning (as the suns rays heat up the debris layer) reaching a peak at around 10:00, and decreasing after this time as the rate of warming decreases. The fact that the flux of \( \partial STOR \) is at a maximum in the morning generates a problem for satellite retrieval of debris thickness using the energy balance method, because the calculation of heat store requires the rate of temperature change, not just instantaneous temperature values.

The average residual of -14.17 is small (Table 4.4), and because \( z_o \) was estimated (0.001 m) and an average value of \( K \) (0.96 Wm\(^{-1}\)K\(^{-1}\)) was applied, the residual value is acceptable, and could be reduced through the tuning of these parameters. Therefore, this highlights that the energy balance model approach has been successful in this application and the assumptions
and simplifications made are justifiable. The successful application of this energy balance model has supported the use of existing energy balance models developed on other debris-covered glaciers. This provides the basis for the generation of an energy balance based model for determining debris thickness on the Miage Glacier, and potential for its application on other debris-covered glaciers.

Figure 5.19: Average hourly summer daytime fluxes 2005 (June-September) measured at the LWS.

To assess the extent to which the energy balance model results from the 01/08/05 follow the ‘typical’ pattern of dominant inputs and output fluxes and conductivity or store values for the 2005 season, the energy balance fluxes were calculated for the entire summer season (Figure 5.19). When compared, the daily 01/08/05 results are similar to the average summer 2005 flux pattern, with the main differences being that the variability in the fluxes throughout the day on the 01/08/05 plot (Figure 5.17), which resulted due to cloud and wind fluctuations on
the energy balance fluxes being smoothed out of the summer average plot (Figure 5.19). The fluxes of \( SHF \) and \( SWR\downarrow \) also reach a greater peak on 01/08/05, with values up to 200 Wm\(^{-2}\) higher than the seasonal averages. Apart from these differences in the amount of \( SWR\downarrow \) received and values of \( SHF \), the general pattern of an increase in the morning and decrease in the evening in all flux is observed.

### 5.7. Extracting debris thickness values through an energy balance model

Both surface temperature and debris thickness are key variables required in an energy balance model to determine sub-debris ablation rates. The development of a model to enable the estimation of these using remote sensing would be an invaluable tool, particularly in inaccessible areas and locations where these variables have not been extensively measured.

#### 5.7.1. Application of the model

Energy balance components were calculated using measurements at the LWS (for 01/08/05 at 10:40) of \( SWR\downarrow \) and \( LWR\uparrow \), air temperature, and wind speed at 2 m (all variables were initially assumed constant across the glacier), to find \( COND \) from Equation 5.3. Surface temperature was provided by an ASTER AST08 surface temperature image for 01/08/05 at 10:40. Standard values of atmospheric constants were used with an initial \( z_o \) of 0.001 m. Because the rate of debris temperature change is unknown when only one satellite image is available, \( \partial STOR \) was estimated based on surface temperature measurements at the LWS for the entire 2005 ablation season. This identified that, on average, 36% of \( COND \) goes into heat store in the debris at the time of image acquisition during fine conditions. Therefore, the resulting \( COND \) flux was multiplied by 0.64 (1 - 0.36). Once all data had been inputted into the model, Equation 4.3 was solved for every pixel to find the conductive heat flux (\( COND \)) as a residual. Next, by assuming that the ice-debris interface is at melting point and that thermal conductivity (\( K \)) is
uniform across the glacier, debris thickness ($d$) can be found as a residual from a re-arranged heat conduction equation (Equation 5.4).

$$K \times \frac{T_s}{\text{COND}} = d \quad (5.4)$$

Where: $K = 0.96$ (Wm$^{-1}$K$^{-1}$), the average value of debris thermal conductivity calculated at 25 point locations (ablation stakes) on Miage Glacier.

The calculated debris thickness at the LWS site was 0.09 m when a $z_o$ of 0.001 m was used, and 0.17 m when the $z_o$ value was increased to 0.01 m, which is close to the measured depth of 0.16 m (although this point measurement and does not necessarily correspond to the 90 x 90 m area sampled by the ASTER pixel). The $z_o$ value used in the model clearly has a large impact upon the resulting debris thickness value. However, little is known about the magnitude of $z_o$ on debris covered glaciers and, therefore, it will be treated as a tuning parameter in this study.

![Graph of estimated debris thickness](image)

**Figure 5.20:** Estimated debris thickness from the energy balance model at the LWS site on 01/08/05 ($z_o = 0.001$ m) for the time period when the assumption of a linear downward temperature profile in the debris is likely to be met (08:00-16:30). Point thickness measured at the LWS site is 0.16 m.
To test the validity of this simplified energy balance model approach to debris thickness estimation, debris thickness was calculated hourly on 01/08/05 at the LWS using measured meteorological variables, and surface temperature values calculated from the measured upwelling longwave radiation flux (from the CRN1 instrument at the LWS). The calculated debris thickness remains relatively constant during the period of the day when the assumption of a linear downward temperature profile in the debris is most likely to be met (08:00-16:30) (Figure 5.20), with the increase around 13:00 associated with cloud cover and probable COND flux divergence in the debris layer. This analysis supports the application of the model at a glacier-wide scale.

![Debris thickness estimates and actual thicknesses at different sites](image)

**Figure 5.21:** Debris thickness estimates and actual thicknesses at different sites, using a fixed air temperature value (measured at the LWS), and measured thermistor surface temperatures.

However, when Equation 5.4 was applied to other locations at similar elevations using a constant value of air temperature, incorrect debris thickness estimates resulted (Figure 5.21). Clearly there was a problem with spatial extrapolation of the energy balance model, with a
negative thickness inferred for one site (site 15). This probably resulted from the incorrect assumption that air temperature depends solely on elevation, whereas air temperature is strongly dependant on surface temperature (itself a function of debris thickness) on a debris-covered glacier through upwelling $LWR\uparrow$ and $SHF$. Consequently, different air temperature values need to be calculated as a function of surface temperature to apply the energy balance debris thickness model across the glacier.

Variations in air temperatures between the upper and lower reaches were investigated through the analysis of lapse rates. When calculated air temperature lapse rates between the upper and lower weather station are plotted, a distinct daily cycle is evident in both daily and seasonal average data (Figure 5.22 a and b). The presence of this daily cycle shows that air temperature is strongly related to surface temperature (Figure 5.22 c), with convective and radiative heating from the debris layer warming air temperatures during the day and sensible heat transfer cooling air temperatures during the night (Brock et al., in press). The large increase in lapse rate between 07:00 and 10:00 occurs due to differential heating of the lower and upper reaches of the glacier, with the lower glacier exposed to solar heating from dawn compared to the upper reaches which is in shade until ~09:00 (Brock et al., in press).

Findings highlighted that air temperature could not be assumed constant across the glacier, due to its variability with elevation and debris thickness. Therefore, a method of air temperature extrapolation was developed. Normally, air temperature is extrapolated across glaciers using a (constant) elevational lapse rate, but on a debris-covered glacier this assumption is invalid due to the highly variable surface temperature. As air temperature was found to be strongly controlled by surface temperature, a regression approach was applied based on measured air and surface temperature at the LWS to obtain an equation that could be applied, based on an assumed relationship of air temperature to surface temperature.
Figure 5.22: a) mean hourly temperature lapse rates between the upper and lower AWS 23/06/07, b) mean daily cycles of hourly temperature lapse rates between the upper and lower AWS for 2006 (LR06) and 2007 (LR07) seasons (Brock et al., in press), c) mean daily cycle of air and surface temperature for the 2005 ablation season, with air temperature measured at 3 heights: 0.5 m, 1 m, and 2 m, surface temperature recorded using surface thermistors (Brock et al., in press).
The regression equation was developed by regressing 2 m air temperature against radiative surface temperature recorded by the CNR1 net radiometer using data recorded at 10 minute intervals on 01/08/05 (Equation 5.5). Temperatures derived from the upwelling $LWR^\uparrow$ flux recorded by the CNR1 net radiometer, using an emissivity of 0.94, were used as this method of surface temperature identification is identical to the way in which surface temperature is derived from an ASTER thermal image. An emissivity of 0.94 was used due to its previous application on the Miage Glacier, and is based on published values for granitic and metamorphic rocks (Brock et al., in press). Similarly, an emissivity value of 0.95 was used by Nicholson and Benn (2006) during a similar study on two debris-covered glaciers supporting the use of 0.94 in this project.

$$T_a = 5.346 + 0.382 \times CRN1 \, T_s$$  \hspace{1cm} (5.5)

**Figure 5.23:** Plot of air vs. surface temperature 01/08/05 to determine whether a relationship existed between the two variables, air temperature vs. CNR1 estimated surface temperature.
Figure 5.23 identifies that a clear and strong linear relationship exists between 2 m air
temperature and surface temperature at this one site on one day, confirmed by a correlation of
0.975 with a p-value of <0.001. However, the relationship is not entirely straightforward over
a 24 hour period as the relationship is cyclical (hysteresis loop). The concentration of data into
two lines represents the morning (cooler) and evening (warmer) temperatures, and highlights
that different relationships exist at these two different times of the day.

Figure 5.24: Surface temperature (CNRI) vs. Air temperature at 2 m 01/08/05, 2 hourly
intervals.

To investigate this further, data were plotted at two hourly timescales (Figure 5.24) so that the
air temperature-surface temperature relationship at the time of the ASTER image could be
investigated and compared to that at other times. This revealed that a strong linear relationship
occurs between 08:00-10:00, when there is a strong link between surface and air temperature
due to convection. A strong relationship is present at this time due to a lull in wind speeds in the early part of the morning under clear sky, calm conditions (Figure 5.24). At other times, however, the relationship was more complicated and variable, due to the variability in wind speed at other times of the day (Figure 5.25).

![Mean wind speeds recorded at the AWS showing a lull in the early morning on both the 01/08/05, and using average values for the summer 2005 period.](image)

Figure 5.25: Mean wind speeds recorded at the AWS showing a lull in the early morning on both the 01/08/05, and using average values for the summer 2005 period.

The relationship of air temperature to surface temperature between 08:00-10:00 (Figure 5.26, Equation 5.6) was, therefore, developed for application in the energy balance debris thickness model.

\[
2 \text{ m air temperature} = 5.785 + 0.3450 \times \text{CNR1 } T_s
\]  

Equation (5.6)

10:00 was used as the upper limit for the regression analysis as the relationship between air temperature and surface temperature begins to increasingly disintegrate after this time. Therefore, data up to 10:40 (time of the ASTER image acquisition) were not included.
However, due to the close proximity in time of the ASTER image to the 10:00 cut off, the regression equation can still be utilised, especially since the strength of the regression relationship equation and resulting air temperature estimates from surface temperature would be lost at other times.

Figure 5.26: Regression plot of air and surface temperature relationship between 08:00-10:00, 01/08/05, CRN1 data relationship

Regression relationships were also developed for other times of the day including 08:00-12:00 and 08:00-14:00. When estimated surface temperature values using these different regression equations were compared to the measured surface temperature values the regression relationship between 08:00-10:00 produced the most similar surface temperature estimates. However, none accounted for the highly variable nature of surface temperature over a 24 hour period.

5.7.2. Testing transferability of air and surface temperature relationship

5.7.2.1. Application to different days with similar weather conditions

To test whether the 08:00-10:00 relationship holds for other days with similar weather conditions a regression equation was developed using 10 days air and surface temperature data.
between 08:00-10:00. Each of the days selected had similar SWR↓ and wind speed values, and the regression coefficient (Figure 5.27) was compared to that generated for 01/08/05 (Figure 5.26). The completion of this comparison is an essential step as it tests the key assumption that the energy balance model is transferable, unlike empirical methods.

**Figure 5.27:** Regression plot of air and surface temperature relationship between 08:00-10:00 on 10 days with similar weather conditions to 01/08/05, both air and surface temperature measured at the LWS.

**Table 5.7:** Air temperature values generated from surface temperature using both regression coefficients from a single day’s data between 08:00-10:00 and 10 days data between 08:00-10:00

<table>
<thead>
<tr>
<th>Surface temperature (°C)</th>
<th>Air temperature (°C) Regression Equation 01/08/05</th>
<th>Regression Equation from 10 days data</th>
</tr>
</thead>
<tbody>
<tr>
<td>22</td>
<td>13.38</td>
<td>14.24</td>
</tr>
<tr>
<td>20</td>
<td>12.69</td>
<td>13.40</td>
</tr>
<tr>
<td>18</td>
<td>12.00</td>
<td>12.56</td>
</tr>
<tr>
<td>16</td>
<td>11.31</td>
<td>11.71</td>
</tr>
<tr>
<td>14</td>
<td>10.62</td>
<td>10.87</td>
</tr>
<tr>
<td>12</td>
<td>9.93</td>
<td>10.02</td>
</tr>
<tr>
<td>10</td>
<td>9.24</td>
<td>9.18</td>
</tr>
<tr>
<td>8</td>
<td>8.55</td>
<td>8.33</td>
</tr>
<tr>
<td>6</td>
<td>7.86</td>
<td>7.50</td>
</tr>
</tbody>
</table>
When the regression coefficients are compared, a relatively similar value is obtained, with the increased variation in surface and air temperatures experienced over the 10 different days accounting for the variability. Because a relatively similar coefficient is obtained for both a single days data (0.345x + 5.7854) and 10 days data (0.4219x + 4.9586) between 08:00-10:00, the transferability of this relationship on days experiencing similar weather conditions is highlighted. This is supported by Table 5.7, where the same or very similar air temperature values are obtained using both regression equations.

**5.7.2.2. Application to different locations on the glacier surface**

Because the surface temperature-air temperature regression equation was developed at a single location (the LWS) for a single year and day (01/08/05), the relationship between air and surface temperature was tested at different locations using field measurements with a portable AWS during 2007. The equipment used, and location of measurements, are identified in section 4.4.5 with measurements taken at a variety of altitudes, both at different elevations and measured transects on the glaciers surface.

From the results obtained, it was clear that the relationship between air and surface temperature at any one point in time across the glacier is not as strong as that between a day’s measured air and surface temperature at the LWS (Figure 5.28). Unfortunately, these data were collected before it was known that 08:00-10:00 was the most useful for developing an air temperature-surface temperature relationship, with measurements made at a later time (ranging from 11:00-14:00) when wind speed (horizontal mixing) was a factor. Hence, these data are of limited use in validating Equation 5.6 (applied in the energy balance debris thickness model) but do highlight the complex relationship of air temperature to surface temperature on debris-covered glaciers.
Figure 5.28: 2m air temperature vs. surface temperature on two different days, 31/08/07 (11:19-12:16) and 03/09/07 (11:30-13:50), with different surface materials highlighted.

5.8. Debris thickness maps derived from 2005 ASTER thermal imagery

Once $T_a$ had been extrapolated to each ASTER pixel using Equation 5.6, debris thickness was estimated for all pixels on the glacier, using the energy balance debris thickness model with $z_o$ values ranging from 0.001 – 0.01 m. It was found that $z_o$ values of 0.002 m and 0.003 m produced the best agreement with field measurements of debris thickness, with a $z_o$ of 0.003 m identifying thicker debris thicknesses both at the snout (0.70 m+) and on the medial moraines (0.50 - 0.60 m) where greater depths are expected (Figure 5.29 c and d). $z_o$ values of <0.002 m and >0.005 m under- and overestimated debris thickness, respectively. These outputs (Figure 5.29 c and d) were then compared to those from the empirical approaches (Equations 5.1 and 5.2, Figure 5.29 a and b).
Figure 5.29: Estimated debris thickness for the entire debris covered portion of the glacier using a) linear empirical approach, b) quadratic empirical relationship, c) energy balance model approach $z_o$ of 0.002 m applied, and $T_a$ equation from 01/08/05 08:00-10:00, d) energy balance model approach $z_o$ of 0.003 m applied, and $T_a$ equation from 01/08/05 08:00-10:00.
A $z_o$ of 0.016 m was also applied, which was calculated at the LWS in 2005 using wind and temperature profile data (Brock et al., in press). However, this resulted in unrealistic debris thickness values (with some minus thicknesses produced). A reason for this was identified as being a problem with the $SHF$ calculations and the use of estimated air temperature values. This occurs as the increase in $z_o$ made the $SHF$ values become very sensitive to input surface and air temperature (and the resulting temperature gradient between air and surface temperature). At some locations an unrealistically large temperature gradient between air and surface temperature is present, and when combined with low wind speeds (Figure 5.30), unrealistic $SHF$ value estimations result. This indicates that at some pixels the estimated air temperatures may not be representative of the pixel, causing a large temperature gradient between the air temperature and surface temperature, which affects the calculated $SHF$ values and leads to inaccurate debris thickness estimates.

![Figure 5.30: Impact of $z_o$, wind speed and temperature gradient on calculated $SHF$](image)

Figure 5.30: Impact of $z_o$, wind speed and temperature gradient on calculated $SHF$
5.9. Evaluation of calculated debris thickness

5.9.1. Visual comparison

When the debris thickness maps from the energy balance approach (Figure 5.29 c and d) and the empirical approaches (Figure 5.29 a and b) are compared, one key difference (and resulting limitation of the two empirical equations) is evident on the empirical approach debris thickness maps. This is the clear underestimation of both debris thickness and its spatial variability, especially in the middle reaches where thicknesses ranging 0.2-0.5 m (and greater) have been measured in the field, and also at the snout which is known to be >1 m from field observations.

The energy balance approach (Figure 5.29 c and d) is more representative of debris thickness on and around the medial moraines and identifies varied debris thicknesses both on (ranging 0.2-0.5 m) and between the medial moraines (ranging 0.10-0.20 m). In addition to this, thick debris (>0.5 m) at the snout, very thin debris (<0.10 m) in the upper reaches are also correctly identified by the energy balance debris thickness model. Overall, therefore, it is clear that the energy balance debris thickness model with a $z_o$ of 0.003 m (Figure 5.29 d) produces the most visually representative debris thickness map, because the debris thickness map using a $z_o$ of 0.002 m (along with the two empirical approaches) (Figure 5.29 a to c) does not highlight as much variability in debris thickness, similarly to the two empirical approaches.

Results were also compared to a previously collected map of debris thickness from 1997 (Figure 5.31). Debris thicknesses estimates from the energy balance approach show broad agreement with the map of field-measured debris thickness, with results most similar when a $z_o$ of 0.003 m was used. However, movement of the debris in the intervening 8 years since 1997 due to ice flow invalidates a detailed comparison. When this map of field based debris thickness measurements was compared to the two empirical approaches the 1997 debris
thickness highlighted that both have underestimated debris thicknesses, especially at the snout and medial moraines.

5.9.2. Numerical comparison

To assess the accuracy of each of the approaches, measured thicknesses at each of the stake locations in 2005, 2006, and 2007 were compared to the estimated debris thickness for the same location using surface temperatures from the ASTER image. Also a correlation analysis was performed between the energy balance debris thicknesses and the empirical approaches. This is followed by a comparison of the total calculated supraglacial debris load for each method. Finally, results from the energy balance approach are compared to an empirical approach previously applied by Mihalcea et al. (2008).

5.9.2.1 Field data – 2005 stake data

Table 4.6 highlights that the greatest range of debris thickness values was obtained with the energy balance debris thickness model, although some overestimates have resulted especially
when a $z_o$ of 0.003 m was applied. The empirical approaches performed best in areas of thinner debris, and underestimated at those stake locations where thick debris (>0.25 m) was recorded. However, since the empirical relationships were developed using the same data to which they are being compared to it is inevitable that they will have an average difference of 0.

Therefore, the field measurements are not an independent test of the debris thickness results from the empirical approaches. Also, another limitation of using the field data for comparison to both methods is that the thickness measurements themselves have a bias towards thin debris due to the difficulties of digging in thick debris, meaning no thicknesses beneath large boulders (which would make up some of the thicknesses in every ASTER pixel) were ever recorded. Hence, if the debris thickness model is working well it would be expected to overestimate the debris thicknesses from the field measurements, which it clearly does.

The spatial resolution of the ASTER imagery provides an explanation for the variance between the measured and estimated thicknesses, which at 90 x 90 m provides an average value for an 8100 m$^2$ area. Also, because the measurements of thickness are single point measurements with a bias towards thinner depths they are unlikely to be representative of the 90 x 90 m area. Therefore, thicknesses from the depth transects taken in 2006 and 2007 will be compared to the estimated depths at the same location, since it involved the systematic sampling of thicknesses in areas of the glacier and are much more suited to verifying ASTER results.
Table 5.8: Differences of estimates from measured thicknesses (at each of the stake locations in 2005) for each of the four images

<table>
<thead>
<tr>
<th>Stake</th>
<th>Measured $d$ (m)</th>
<th>$d$ difference: Energy balance, $z_o$ 0.002 m</th>
<th>$d$ difference: Energy balance, $z_o$ 0.003 m</th>
<th>$d$ difference: Empirical linear (m)</th>
<th>$d$ difference: Empirical quadratic (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>0.08</td>
<td>-0.03</td>
<td>-0.03</td>
<td>0.03</td>
<td>0.00</td>
</tr>
<tr>
<td>2</td>
<td>0.10</td>
<td>-0.04</td>
<td>-0.04</td>
<td>-0.01</td>
<td>0.00</td>
</tr>
<tr>
<td>3</td>
<td>0.04</td>
<td>0.00</td>
<td>0.00</td>
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</tr>
<tr>
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<td>-0.05</td>
<td>-0.05</td>
<td>-0.05</td>
<td>-0.06</td>
</tr>
<tr>
<td>5</td>
<td>0.27</td>
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<td>-0.14</td>
<td>-0.12</td>
<td>-0.16</td>
</tr>
<tr>
<td>6</td>
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<td>0.06</td>
<td>0.07</td>
<td>0.04</td>
</tr>
<tr>
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<td>0.13</td>
<td>0.07</td>
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</tr>
<tr>
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<td>0.10</td>
<td>0.07</td>
<td>0.02</td>
</tr>
<tr>
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<td>0.09</td>
<td>0.04</td>
<td>0.04</td>
</tr>
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<td>0.01</td>
<td>0.02</td>
<td>0.05</td>
<td>0.01</td>
</tr>
<tr>
<td>11</td>
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<td>-0.01</td>
<td>0.05</td>
<td>-0.05</td>
<td>-0.04</td>
</tr>
<tr>
<td>12</td>
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<td>0.15</td>
<td>0.08</td>
<td>0.08</td>
</tr>
<tr>
<td>13</td>
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<td>0.06</td>
<td>0.14</td>
<td>-0.01</td>
<td>0.02</td>
</tr>
<tr>
<td>14</td>
<td>0.16</td>
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<td>0.08</td>
<td>0.02</td>
<td>0.03</td>
</tr>
<tr>
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<td>0.15</td>
<td>0.04</td>
<td>0.07</td>
<td>0.03</td>
<td>0.04</td>
</tr>
<tr>
<td>16</td>
<td>0.11</td>
<td>0.07</td>
<td>0.10</td>
<td>0.06</td>
<td>0.05</td>
</tr>
<tr>
<td>17</td>
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<td>0.00</td>
<td>-0.04</td>
<td>-0.05</td>
</tr>
<tr>
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<td>-0.05</td>
<td>-0.04</td>
<td>-0.07</td>
</tr>
<tr>
<td>19</td>
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<td>-0.02</td>
</tr>
<tr>
<td>20</td>
<td>0.16</td>
<td>0.08</td>
<td>0.15</td>
<td>0.03</td>
<td>0.04</td>
</tr>
<tr>
<td>21</td>
<td>0.27</td>
<td>0.03</td>
<td>0.16</td>
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<td>-0.06</td>
</tr>
<tr>
<td>22</td>
<td>0.12</td>
<td>0.14</td>
<td>0.22</td>
<td>0.06</td>
<td>0.09</td>
</tr>
<tr>
<td>23</td>
<td>0.20</td>
<td>0.06</td>
<td>0.14</td>
<td>-0.01</td>
<td>0.02</td>
</tr>
<tr>
<td>24</td>
<td>0.16</td>
<td>0.27</td>
<td>0.59</td>
<td>0.04</td>
<td>0.12</td>
</tr>
<tr>
<td>25</td>
<td>0.55</td>
<td>-0.11</td>
<td>0.24</td>
<td>-0.35</td>
<td>-0.27</td>
</tr>
<tr>
<td><strong>Average difference</strong></td>
<td><strong>0.03</strong></td>
<td><strong>0.09</strong></td>
<td><strong>0.00</strong></td>
<td><strong>0.00</strong></td>
<td><strong>0.00</strong></td>
</tr>
<tr>
<td><strong>RMSE</strong></td>
<td><strong>0.09</strong></td>
<td><strong>0.16</strong></td>
<td><strong>0.09</strong></td>
<td><strong>0.08</strong></td>
<td><strong>0.08</strong></td>
</tr>
</tbody>
</table>
Due to issues of the ASTER spatial resolution, the transect data available are better suited to verifying the debris thickness estimates, as they provide debris thickness measurements over a greater area compared to a single point measurement and are therefore more representative of an ASTER pixel (90 x 90 m). An average debris thickness value can also be generated using data collected over a transect, providing a better means of comparison to ASTER debris thickness estimates, because debris thickness from an ASTER pixel will be an average thickness over the 90 x 90 m pixel area (Mihalcea et al., 2008).

From Table 5.7 a and b it is clear that the methods which perform best are the energy balance debris thickness model on the lower transects in 2006 and 2007, and the empirical approaches on the middle 2006 and upper 2007 transects. The size of the transects provides a reason for the differences between the measured and estimated debris thickness values, because in 2007 the upper transect consisted of only 14 sample locations over a 45 x 30 m area, and the lower transect in 2006 had only 20 sample points over two 100 m transect arms. Whereas, the upper lower and middle transects in 2006 and lower transect in 2007 comprised of 40 sample points over four 100 m transect arms. Therefore, the 2006 lower transect and 2007 upper transect were not representative of an ASTER pixel, providing an explanation for the difference between the ASTER debris thickness value and the measured average in the field.

Another issue in validating ASTER debris thickness estimates from field based debris transects is the matching of co-ordinates of the transect area to the ASTER pixel locations, since to validate an ASTER debris thickness the transect used should be taken within the corresponding pixel location on the ground. However, this is difficult to achieve due to accessibility issues at some locations on the glacier (due to presence of heavily crevassed regions, ice cliffs and steep sided medial moraines). Therefore, debris transects may span two
(or more) ASTER pixels, meaning they may not be representative of the ASTER debris thickness value requiring validation.

Table 5.9: Average debris thickness for transect area and ASTER debris thickness estimate for the same transect (ASTER thickness estimate for entire transect calculated using surface temperature extracted from the mid point of each transect), where: 1 = Energy balance method \(T_a\) equation 08:00-10:00 \(z_o\) 0.002 m, 2 = Energy balance method \(z_o\) 0.003 m \(T_a\) equation 08:00-10:00, 3 = empirical linear approach, 4 = empirical quadratic approach a) 2006, all transects 5 m spacing, b) 2007, upper transect 5 m and lower transect 20 m spacing a)

<table>
<thead>
<tr>
<th>Sample points</th>
<th>Average transect d (m)</th>
<th>Estimated d (m) 1</th>
<th>Estimated d (m) 2</th>
<th>Estimated d (m) 3</th>
<th>Estimated d (m) 4</th>
</tr>
</thead>
<tbody>
<tr>
<td>Lower transect (100 m² area)</td>
<td>40</td>
<td>0.23</td>
<td>0.17</td>
<td>0.19</td>
<td>0.20</td>
</tr>
<tr>
<td>Middle transect (100m² area)</td>
<td>40</td>
<td>0.32</td>
<td>0.09</td>
<td>0.09</td>
<td>0.23</td>
</tr>
<tr>
<td>Upper transect (100 m line)</td>
<td>10</td>
<td>0.18</td>
<td>0.22</td>
<td>0.27</td>
<td>0.16</td>
</tr>
</tbody>
</table>

b)

<table>
<thead>
<tr>
<th>Sample points</th>
<th>Average transect d (m)</th>
<th>Estimated d (m) 1</th>
<th>Estimated d (m) 2</th>
<th>Estimated d (m) 3</th>
<th>Estimated d (m) 4</th>
</tr>
</thead>
<tbody>
<tr>
<td>Upper transect (45 x 30 m area)</td>
<td>14</td>
<td>0.17</td>
<td>0.06</td>
<td>0.06</td>
<td>0.14</td>
</tr>
<tr>
<td>Lower transect (two 100 m arms)</td>
<td>20</td>
<td>0.2</td>
<td>0.17</td>
<td>0.20</td>
<td>0.21</td>
</tr>
</tbody>
</table>

5.9.2.3. Supraglacial debris load

The total supraglacial debris load was calculated by adding together all of the pixels debris thickness estimates, next this value was adjusted for assumed void space. Void space was measured in the field in 2005, where part of the debris surface was removed and placed into a bucket. The volume of this debris was then measured, next the amount of water required to fill the bucket to the level of the debris was recorded (Maconachie, 2005). This method assumes that the water is filling the void spaces, and the volume of water required to fill these void spaces can be compared to the debris volume to calculate that percentage of void space. The
average value of void space measured at the Miage Glacier was 38%. Therefore, the total debris load calculated was multiplied by 0.62 (1-0.38) to account for void space.

Table 5.10: *Total supraglacial debris load, adjusted for assumed void space (average glacier value of 38% applied)*

<table>
<thead>
<tr>
<th>Method</th>
<th>Total supraglacial debris load (adjusted for assumed void space) m³</th>
</tr>
</thead>
<tbody>
<tr>
<td>Empirical approach - Linear equation</td>
<td>35.39</td>
</tr>
<tr>
<td>Empirical approach - quadratic equation</td>
<td>37.25</td>
</tr>
<tr>
<td>Energy balance model - ( z_o ) 0.002 m</td>
<td>40.69</td>
</tr>
<tr>
<td>Energy balance model - ( z_o ) 0.003 m</td>
<td>51.26</td>
</tr>
</tbody>
</table>

Table 5.10 identifies that the greatest debris loads are identified by the two energy balance approaches, whereas the empirical approaches have lower debris load estimates due to the underestimation of both these methods in areas of thicker debris. The method with the higher debris load is the energy balance approach using a \( z_o \) of 0.003 m, this method also performed most successfully in the visual comparison. Confirming that it produced the most reliable debris thickness estimates.

5.9.2.4. *Comparison of Mihalcea et al., (2008) thickness estimates*

As mentioned previously, Mihalcea et al., (2008) completed a study of debris thickness estimation on the Miage Glacier using a more detailed empirical approach. Due to the limitations with the linear approach (utilising a single empirical equation) for correctly identifying thickness across the entire glacier, 8 different regression equations were generated. These were based on the different relationship of debris thickness and surface temperature at different elevations on the glacier, with a different equation for every 100 m elevation band (Equations 5.7 to 4.14). The results from this are compared to those obtained using the energy balance approach in this study.
\begin{align*}
1720–1800 \ m : d &= 0.047 \times T_s - 0.94 \quad (5.7) \\
1801–1900 \ m : d &= 0.033 \times T_s - 0.63 \quad (5.8) \\
1901–2000 \ m : d &= 0.030 \times T_s - 0.67 \quad (5.9) \\
2001–2100 \ m : d &= 0.020 \times T_s - 0.3 \quad (5.10) \\
2101–2200 \ m : d &= 0.044 \times T_s - 1.0 \quad (5.11) \\
2201–2300 \ m : d &= 0.016 \times T_s - 0.16 \quad (5.12) \\
2301–2400 \ m : d &= 0.015 \times T_s - 0.14 \quad (5.13) \\
2401–2500 \ m : d &= 0.010 \times T_s - 0.06 \quad (5.14)
\end{align*}

When the debris thickness map is compared to the energy balance approach it is clear that the debris thicknesses obtained using a \( z_o \) of 0.002 m (Figure 5.32 a) are most similar to the Mihalcea et al., (2008) results (Figure 5.32 c). Compared to the energy balance approach with \( z_o \) 0.003 m map (Figure 5.32 b), which identifies areas of thicker debris in the middle reaches of the glacier of 0.4 – 0.5 m. However, these thicker areas do correspond to the locations of the medial moraines on the glacier, and are supported by values measured in the field with thicknesses reaching 0.4 – 0.5 m (Figure 5.32 b).

The empirical approach of Mihalcea et al., (2008) does identify more variability in debris thickness towards the upper reaches of the glacier with thicknesses ranging 0.05-0.2 m identified, unlike the energy balance approaches which identify much more extensive areas of very thin debris in this region (0.005-0.10 m). Despite these differences in the middle and upper reaches, the thicknesses estimated at the snout of the glacier are very similar in all approaches (although the model with a \( z_o \) of 0.003 m identifies areas of debris thicker than 0.5 m) with the thickest debris at the two lobes mixed in with patches of debris varying in thickness (from 0.10 – 0.30 m). These results highlight that since the Mihalcea et al., (2008) thickness map was generated from extensive field data and detailed, highly calibrated
empirical relationships, the close correspondence with the energy balance debris thickness model results provides strong support for this method and its application.

Figure 5.32: Comparison of thickness estimates obtained using an energy balance approach $z_o$ 0.002 m, air temperature regression equation 01/08/05 08:00-10:00, b) energy balance approach $z_o$ 0.00 m3, air temperature regression equation 01/08/05 08:00-10:00, c) a empirical linear approach with different equations for each 100 m elevation band (starting at 1700-1800, up to 2400 – 2500 m) (Mihalcea et al., 2008) $T_s$ from 01/08/05 ASTER image.
Comparison of the total supraglacial debris loads calculated by both the energy balance approach and Mihalcea et al., (2008) could not be completed as the actual values of debris thickness for each of the pixels in the Mihalcea et al., (2008) method were not available. Another consequence of not having the actual debris thickness estimates for the Mihalcea et al., (2008) method is that a difference map could also not be produced.

5.10. Sensitivity analysis

After its initial application to the 2005 ASTER image a number of sensitivity tests were carried out on the model to assess the impact of changing certain parameters upon the debris thickness values calculated. Sensitivity of debris thickness to variation in: conductivity, emissivity, $z_o$, slope and aspect (assumed constant values of zero in the initial application in $SWR^\uparrow$ and $LWR^\downarrow$) were analysed.

5.10.1. Conductivity

Initially an average conductivity value of 0.96 Wm$^{-1}$ K$^{-1}$ was used. However, because this value will vary at different locations on the glacier due to variations in debris thickness and rock type, three different conductivities were applied.

<table>
<thead>
<tr>
<th>Conductivity Value</th>
<th>Mean debris thickness (m)</th>
<th>Standard deviation</th>
<th>LWS debris thickness estimate</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.96 (glacier average)</td>
<td>0.14</td>
<td>0.05</td>
<td>0.09</td>
</tr>
<tr>
<td>1.36 (stake 7)</td>
<td>0.20</td>
<td>0.08</td>
<td>0.13</td>
</tr>
<tr>
<td>0.50 (stake 12)</td>
<td>0.07</td>
<td>0.03</td>
<td>0.05</td>
</tr>
</tbody>
</table>
Results (Table 5.8) show that changing the conductivity value to the maximum and minimum conductivity values recorded on the Miage Glacier in 2005 had an impact upon the debris thickness calculated. An estimated debris thickness is greater when a higher conductivity value is used, because the debris estimation equation multiplies the conductivity value by the surface temperature then divides it by $\text{COND}$.

### 5.10.2. Surface roughness value ($z_o$)

A number of different $z_o$ values were applied to this model, due to the variability of $z_o$ across a glacier due to changes in the dimension, form and density of surface roughness elements (Oke, 1987; Brock et al., 2006), with values ranging 0.001 m on smooth glacier ice surface to 0.05 m on very rough glacier ice (Brock et al., 2006). As a result the $z_o$ value will increase with increasing height, surface area, and density of debris on the glaciers surface (Brock et al., 2006). Values applied consisted of 0.001 m, 0.002 m and 0.003 m, and a $z_o$ value of 0.016 m calculated for the Miage Glacier at the LWS in 2005 using wind and temperature profile data (Brock, et. al. in press). The debris thickness values obtained varied significantly with a change in $z_o$, and show that the debris thickness initially increases with an increase in $z_o$ (Table 5.9).

<table>
<thead>
<tr>
<th>$z_o$ Value</th>
<th>Mean debris thickness (m) (using all ASTER pixels on the glacier)</th>
<th>Standard deviation</th>
<th>LWS Debris thickness estimate (actual depth 0.16m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.001</td>
<td>$0.14$</td>
<td>$0.05$</td>
<td>$0.09$</td>
</tr>
<tr>
<td>0.002</td>
<td>$0.18$</td>
<td>$0.08$</td>
<td>$0.10$</td>
</tr>
<tr>
<td>0.003</td>
<td>$0.22$</td>
<td>$0.14$</td>
<td>$0.11$</td>
</tr>
<tr>
<td>0.016</td>
<td>$-0.14$</td>
<td>$1.89$</td>
<td>$0.23$</td>
</tr>
<tr>
<td>0.010</td>
<td>$-1.94$</td>
<td>$12.11$</td>
<td>$0.17$</td>
</tr>
</tbody>
</table>

As identified in section 5.7 results also indicated that when the $z_o$ is raised to 0.016 m (value measured at the LWS site) the $\text{SHF}$ value calculated becomes very sensitive to errors in input
air temperature (due to its estimation from surface temperature). This, in turn, has a significant impact upon the debris thickness values calculated (Table 5.9). Air temperature values are only known accurately at the LWS site highlighting an extrapolation problem to other locations on the glaciers surface, due to the impact of incorrect air temperature estimates on the \( SHF \) values calculated, which become unrealistically high or low when the \( z_o \) of 0.016 m is used, due to the resulting high temperature gradient and impact of low wind speeds. Such conditions are common around the time of image acquisition (10:40). Therefore, lower \( z_o \) values will be retained in this model – which still provides accurate debris thickness estimates, with a focus for future work on the further investigation of the calculation of \( SHF \) fluxes to remove the problem of over and under estimation.

5.10.3. Emissivity

The impact of different emissivity values on the calculation of the surface temperature from the CNR1 temperature data also needs to be explored. As the presence of different rock types on a glacier’s surface will have an impact upon the use of a single emissivity value, due to the impact of different colours, surface roughness, and moisture content.

Figure 5.33 shows that a change in emissivity value has a significant impact upon the surface temperature recorded by the CNR1 sensor (Table 4.10), with an increase of just 0.02 resulting in a change in temperature of a few degrees. Overall, if the emissivity is increased, the temperature decreases, and when it is decreased, the temperature increases (Figure 5.33). This highlights the importance of selecting the correct emissivity, due to its influence upon surface temperature and its resulting effect upon debris thickness estimates using the developed model (Table 5.10).
Figure 5.33: CRNI $T_s$ values with different emissivity values applied.

Table 5.13: Impact of different emissivity values on surface temperatures (10:40 01/08/05) and resulting debris thickness estimates at the LWS (measured depth 0.16 m) using the energy balance model (air temperature measured), $z_0$ 0.001 m, conductivity 0.96 Wm$^{-1}$ K$^{-1}$

<table>
<thead>
<tr>
<th>Emissivity value</th>
<th>Surface temperature ($^\circ$C) at 10:40 01/08/05</th>
<th>Debris thickness at LWS (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.96</td>
<td>23.20</td>
<td>0.10</td>
</tr>
<tr>
<td>0.94</td>
<td>24.76</td>
<td>0.13</td>
</tr>
<tr>
<td>0.93</td>
<td>25.56</td>
<td>0.15</td>
</tr>
<tr>
<td>0.92</td>
<td>26.37</td>
<td>0.17</td>
</tr>
<tr>
<td>0.90</td>
<td>28.02</td>
<td>0.24</td>
</tr>
</tbody>
</table>

Therefore, it was important to check the emissivity used in the atmospheric correction of the ASTER image. However, as mentioned in chapter 3, the emissivity used by the ASTER team is an average of that recorded by each pixel. Therefore, the value being used should be representative of the emissivity present within the average 90 x 90 m pixel area of the study site in question. Ideally, a different emissivity value would be used for each change in rock type/land cover in each pixel, however, for that a map of debris cover type would be required. This will be investigated in chapter 8.
5.10.4. Slope and aspect

The model was initially applied to all ASTER pixels on the glacier with constant slope and aspect (assuming a flat horizontal glacier). Slope and aspect values were incorporated in the calculation of \(SWR_{\downarrow}\) at each pixel (using the spreadsheet model of Brock and Arnold (2000)), which was previously assumed constant across the glacier (with a measured value at the LWS at 10:40 01/08/05 used in the initial model). Sensitivity testing focused firstly upon using the average slope (14°) and aspect (135°) values for the entire glacier and, secondly, using the individual slope and aspect value from each pixel, obtained from the 1991 orthophoto DEM (details of its generation and accuracy are found in section 4.5.2) (Table 5.14).

<table>
<thead>
<tr>
<th>Slope/aspect</th>
<th>Mean debris thickness (m)</th>
<th>Standard deviation</th>
<th>LWS Debris thickness estimate (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Whole glacier average</td>
<td>0.23</td>
<td>0.52</td>
<td>0.15</td>
</tr>
<tr>
<td>Single pixel value</td>
<td>0.30</td>
<td>0.18</td>
<td>0.11</td>
</tr>
</tbody>
</table>

This DEM was re-sampled from a 10 m to a 90 m spatial resolution, so it was the same spatial resolution as the ASTER thermal image (90 x 90 m) to enable analysis during this study. The orthophoto DEM (1991) was selected for use over a more recent ASTER DEM (2000, 2004, 2005, and 2006 available) due to the poor vertical accuracy of ASTER DEMs in complex mountainous terrain (highlighted in chapter 7), which would have a significant impact upon the slope and aspect values calculated.
Figure 5.34: Debris thickness estimates when a different aspect/slope values are used for each pixel, b) average slope and aspect values are used in the model – both obtained from an orthophoto DEM (1991), with a $z_o$ of 0.001 m.

Table 5.15: Total supraglacial debris load, adjusted for assumed void space (average glacier value of 38% applied)

<table>
<thead>
<tr>
<th>Method</th>
<th>Total supraglacial debris load (adjusted for assumed void space) m$^3$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Energy balance approach – $SWR\downarrow$ calculated using each pixels slope and aspect value</td>
<td>53.52</td>
</tr>
<tr>
<td>Energy balance approach – $SWR\downarrow$ calculated using average slope and aspect value</td>
<td>67.62</td>
</tr>
</tbody>
</table>

Initially, it was anticipated that the inclusion of each pixel’s slope and aspect value would increase the accuracy of the results from the debris thickness model, and perform better than when the average slope and aspect value was used. However, Figure 5.34 and Table 5.15 highlights that the use of average slope and aspect values appears to have been most successful, as it clearly identifies the medial moraines in the middle reaches, and the increase in debris thickness with distance down glacier. The estimated debris thickness at the LWS was closer to the measured depth (0.16 m) when the average slope/aspect value for the glacier is used, rather than its corresponding pixel (Table 5.14). Table 5.15 also supports the use of an
average slope and aspect value, as a greater total supraglacial debris load is estimated when an average slope and aspect value is used.

The results using each pixel’s individual slope and aspect tends to underestimate debris thickness, with thicknesses of only 0.2 and 0.3 m identified in the middle reaches and 0.3 – 0.4 m at the snout. Also, Figure 5.33 a, highlights that negative debris thickness values are estimated at 18 pixel locations, and debris thicknesses greater than 2 m at 3 pixel locations, which is unrealistic, and indicates a clear problem with the debris thickness calculation at these locations. This is the result of the estimated air temperature values applied at these pixels, and their resulting impact on SHF values and debris thickness estimates (highlighted previously in section 5.8), which consequently restricted the $z_0$ value to 0.001 m as anything greater increased the number unrealistic debris thickness estimates.

A reason for the differences between the debris thickness values calculated using the average slope and aspect values and individual pixel values may be a result of the age of the orthophoto DEM (1991), and that debris redistribution and glacier thinning will have altered the slope and aspect values in some locations (especially on or around the steeper moraines and glacier snout), which in turn will impact the calculated $SWR\downarrow$, and the debris thickness estimates. The age difference between the orthophoto (1991) and date of study (01/08/05) will have had a greater impact on individual slope and aspect values compared to the average glacier slope and aspect value.

5.10.5. Application of variable LWR values

The $LWR\uparrow$ used in the model was recorded at the LWS at 10:40 on 01/08/05 and was assumed constant across the glacier. However, $LWR\uparrow$ will have varied across the glacier due to changes in debris thickness, as it will be larger (more negative) on thicker warm debris, and smaller on
thinner cold debris. Therefore, $LWR^\uparrow$ was estimated based on the average debris emissivity calculated for the glacier, surface temperature (in degrees Kelvin) from the ASTER image (Equation 5.15), and Stefan Boltzman constant (energy radiated per unit of surface area). For $LWR^\downarrow$ the measured value at the LWS was used.

$$LWR^\uparrow = 0.94 * 0.0000000567 * T_s^4$$  \hspace{1cm} (5.15)

*Where: 0.94 = average emissivity calculated for the glacier*

The estimated $LWR^\uparrow$ (Figure 5.34 a) generally reflects the debris thickness distribution, as the emitted $LWR^\uparrow$ is a function of the debris thickness. Therefore, those areas with thinner debris such as the upper reaches will have much lower $LWR^\uparrow$ values compared to the thicker debris at the snout. Results show (Figure 5.35 b and c) that the inclusion of $LWR^\uparrow$ for each pixel tends to underestimate debris thicknesses throughout the glacier using $z_o$ values of 0.002 m and 0.003 m. These debris thickness estimates are still underestimated even when the measured $z_o$ (calculated at the LWS) of 0.016 m was used, with no thickness values greater than 0.20 m generated (Figure 5.35 b). Once the $z_o$ value was increased to 0.03 m, the values became more representative (Figure 5.35 c).

When the calculated $SWR^\downarrow$ (using each pixels slope and aspect value) are also included, the calculated thickness values are much more varied (Figure 5.35 d, Table 5.16). However, due to problems with the air temperatures in $SHF$ calculation (discussed previously), some negative debris thickness values have resulted (when a $z_o$ of 0.016 m or greater is used). This problem only occurs at 4 pixels, all of which are marginally located at the snout and where the chance of a large slope and aspect error is greatest. As these occur in areas where debris is known (from fieldwork) to be very thick (>1 m) they can, therefore, either be excluded from analysis or changed manually.
Figure 5.35: a) Calculated LWR↑ values, b) debris thickness when LWR↑ is calculated for each ASTER pixel, using a z₀ of 0.016 m, c) debris thickness when LWR↑ is calculated for each ASTER pixel, using a z₀ of 0.03 m, d) debris thickness when calculated SWR↓ from each pixels slope and aspect is used with the calculated LWR↑ values, using a z₀ of 0.016 m.
Table 5.16: Total supraglacial debris load, adjusted for assumed void space (average glacier value of 38% applied)

<table>
<thead>
<tr>
<th>Method</th>
<th>Total supraglacial debris load (adjusted for assumed void space) m³</th>
</tr>
</thead>
<tbody>
<tr>
<td>Energy balance approach – $LWR\uparrow$ calculated, $z_0\ 0.016$ m</td>
<td>16.87</td>
</tr>
<tr>
<td>Energy balance approach – $LWR\uparrow$ calculated, $z_0\ 0.03$ m</td>
<td>30.14</td>
</tr>
<tr>
<td>Energy balance approach – $LWR\uparrow$ calculated, and $SWR\downarrow$ calculated using each pixels slope and aspect $z_0\ 0.03$ m</td>
<td>21.18</td>
</tr>
</tbody>
</table>

The distribution of debris thickness identified by the model using both estimated $SWR\downarrow$ and $LWR\uparrow$ (Figure 5.34 d) is similar to the ‘simplified’ model results (Figure 5.29 c and d).

However, the amount of detail in debris thickness identified in this more physically-based model is not as varied as the ‘simplified’ model. Also the total supraglacial debris load estimated is much lower when the $LWR\uparrow$ and $SWR\downarrow$ values are calculated for each pixel (21.18 m³, Table 5.16) compared to the value obtained using the ‘simplified’ model (40.69 m³ or 51.26 m³, table 5.10). This highlights that the estimated debris thicknesses when both $LWR\uparrow$ and $SWR\downarrow$ values are calculated for each pixel are too thin.

One reason for this, is the inclusion of the estimated $SWR\downarrow$, which is calculated from a DEM from a much earlier date, and the slope and aspects will have changed in the pixels since the DEM acquisition (1991). Also, due to the varied and undulating terrain of the Miage Glacier, the within pixel variability of slope and angle will be significant, meaning the slope/aspect value for each pixel is an average for the 10 x 10 m area highlighting limitations of using a 10 m DEM.

The inclusion of calculated $LWR\uparrow$ and $SWR\downarrow$ for each pixel into the previous ‘simplified’ model has identified that the more physically-based the model becomes, the more it is affected by quality of data used. For example, the spatial resolution and age of the DEM has a significant impact, as when an older or coarser DEM is applied the likelihood of slope and
aspect errors increases. This highlights the requirements for DEMs acquired near the same
time of the surface temperature image so that slope and aspect values will not have varied
significantly. ASTER DEMs could be generated from the same image that the surface
temperature image is developed. However, the limitations of ASTER DEMs in this terrain
(identified in chapter 6), and their spatial resolution (30 m) for slope and aspect identification
limits their applicability.

Despite the loss of detail in debris thickness variability, and debris thickness underestimates
with the inclusion of calculated $SWR_\downarrow$ and $LWR_\uparrow$, the factors causing this are identified, and
solutions (good quality data, from the same time as the surface temperature image) suggested.
This increases the support for the models transferability, because only limited ‘tuning’ of the
model is required through the $z_o$ value as a tuning parameter (discussed previously in section
5.7.1). Further investigation into the development of the ‘simplified’ model to a more
physically-based model is still needed, as the air temperature regression equation is set as the
same for the lower and upper glacier, despite the differences in this relationship that will result
at different elevations. Also, despite $LWR_\uparrow$ being calculated the model at present doesn’t
account for the variation of $SWR_\downarrow$ and $LWR_\downarrow$ at each pixel and this, therefore, requires further
analysis.

5.11. Comparison with 2004 debris thickness map – from ASTER

To test the transferability of the energy balance debris thickness model it was applied to a
different ASTER image from 2004. As there was only 1 year between the two images, debris
thickness maps derived from the ASTER temperature data should be more or less identical.
Successful application of this would support the application of the model at other dates and
highlight the potential for development for application on other glaciers. An ASTER image
from 2004 (29th July) was selected for use as it was the best image available from the ASTER
archive that was both snow and cloud free, and was nearest in the time of year to the original test image on 01/08/05.

AWS data from the Miage Glacier surface was not available for this year. Therefore, the surface temperatures of the two ASTER AST08 images were compared (Figure 5.36), and it was clear that the surface temperature on both of these dates was relatively similar. This means that the air temperature equation developed for the 01/08/05 between 08:00-10:00 could be applied. Measured $LWR_\downarrow$, wind speed and air vapour pressure was also not available for this date (no AWS on the glacier in 2004). Therefore, the same data from 01/08/05 were used in the model for its initial test, to see whether its transfer to another date with this ‘simplified model’ would work. $SWR_\uparrow$ was calculated in a spreadsheet model of Brock and Arnold (2000), using the parameters of latitude (7), longitude (15), day (Julian day 211), time (11:00), albedo (0.13), and $LWR_\uparrow$ was calculated using Equation 5.15. Once these data were put in to the model, the resulting calculated debris thickness (Figure 5.37 a and b) were compared to the

Figure 5.36: Surface temperatures in a) 29/07/04, b) 01/08/05 from corresponding AST08 images.
previous thicknesses obtained using the ‘simplified’ model for 2005 ASTER data (Figure 4.30 c and d).

**Figure 5.37:** 27/07/04 Debris thickness with air temperature calculated using 08:00-10:00 equation, a) $z_0$ 0.002 m, b) $z_0$ 0.003 m.

**Figure 5.38:** Differences between debris thickness estimates in 2005 and 2004, a) $z_0$ 0.002 m b) $z_0$ 0.003 m.
Table 5.17: Total supraglacial debris load, adjusted for assumed void space (average glacier value of 38% applied)

<table>
<thead>
<tr>
<th>Method</th>
<th>Total supraglacial debris load (adjusted for assumed void space) m³</th>
</tr>
</thead>
<tbody>
<tr>
<td>Energy balance model - (z_o) 0.002 m</td>
<td>37.07</td>
</tr>
<tr>
<td>Energy balance model - (z_o) 0.003 m</td>
<td>45.14</td>
</tr>
</tbody>
</table>

When the 2004 and 2005 debris thickness maps (Figure 5.29 c and d, 5.37 a and b) are compared visually, some similarities are apparent where thin debris is dominant in the upper reaches. However, a number of differences are also evident (Figure 5.38), with the main differences occurring in the middle reaches around the upper reaches, medial moraines and at the snout. This may be explained by the movement of debris in these regions of undulating topography – especially the areas surrounding the medial moraines, where temperatures are relatively similar between the two dates (Figure 5.37 a and b), leading to changes of around 5-10 cm.

The movement of debris between the two dates provides an explanation for the presence of larger changes in debris thickness (up to 20 cm on the moraines, and up to 50 m at the snout) identified between 29/07/04 and 01/08/05. Slight differences in surface temperature between the two dates also provides an explanation for the differences experienced in some locations, with areas on the snout being warmer (by ~ 5 °C) in 2004 than in 2005. Smaller differences in debris thickness between 2004-2005 occur on the images with a \(z_o\) of 0.002 m. However, the 2004 debris map with a \(z_o\) of 0.003 m produced a more detailed debris thickness map (Figure 5.36 b) with detail of the medial moraines and thicker debris at the snout (both of which are not highlighted in the 2004 with a \(z_o\) of 0.002 m).

When the total supraglacial debris load is compared between the two images large differences are apparent (Table 5.10, 5.17), with the 2004 debris thickness map producing lower
supraglacial load estimates. This highlights that the 2004 debris map underestimates debris thicknesses slightly, and may be resolved through the application and testing of different $z_o$ values. Another explanation is the use of data from 2005 (measured $LWR$, wind speed, air vapour pressure, and air temperature equation) in estimating thicknesses for 2004, the use of data closer to the date of image acquisition in 2004 should produce more accurate debris thickness estimates. To resolve this data collected from the same year should be utilised, when data is not available for the glacier alternative sources of this data should be sought, including local weather stations or NCEP/NCAR (National Centres for Environmental Prediction-National Centre for Atmospheric Research) reanalysis data, which can be obtained globally for different elevations from NOAA (National Oceanic and Atmospheric Administration - air temperature and wind speed).

The study of debris thickness variations over time is important from a glaciological perspective, due to the different response of debris-covered glaciers to climatic changes, and the differential impact of a debris layer depending upon its thickness. A thick debris layer insulates and protects the ice beneath from melt, and debris layers are thickest at the glaciers snout, which results in widespread thinning of a glaciers surface rather than frontal recession during periods of climatic warming (Diolaiuti et al., 2003). However the opposite is encountered and ablation enhanced if an increase in thin patchy debris occurs. Therefore, the monitoring of any changes in debris thickness on a debris-covered glacier is of key importance, as any changes will have a significant impact upon the glacier ice beneath. This in turn, has an impact upon melt rates (and melt rate predictions) from these glaciers and affects those who are dependant upon the glacier as a freshwater resource. Also under negative mass balance conditions debris cover will increase on a glacier due to the reduction in glacier transport rates (Kirkbride, in press) and melt out of englacial sediments (Diolaiuti et al., 2009).
Compared to a reduction in debris cover under positive mass balance conditions as debris transport becomes more effective and melt out of englacial sediments does not occur.


As well as testing the transferability of the energy balance method, the Mihalcea et al., (2008) method was also applied to the 2004 ASTER image. To achieve this, extracted surface temperature data from 2004 were sorted according to elevation (AST08 2004 image was stacked with the 1991 orthophoto DEM, meaning surface temperature and elevation of each pixel could be extracted in an ASCII file), and the 8 different equations applied to the ASTER data according to the pixels elevation.

As these two maps are only a year apart, the differences in debris thickness between them should be minimal. Both maps (Figure 5.39 a and b) indicate that a majority of the glacier tongue has debris around 20-30 cm in thickness and that thinner debris is found in the upper reaches. The main difference is that the 2004 map does not highlight as much detail/extent of the medial moraines, also, that thinner debris thicknesses extend further down the main glacier tongue, and thinner debris on the southern lobe compared to 2005.

The main problem with debris thickness values obtained from the 2004 image using the Mihalcea et al., (2008) method is that it identifies negative debris thicknesses. These result due to the empirical equations used, which were developed based on the relationship between air temperature and surface temperature at a certain time on a certain day. Therefore, the results demonstrate that the Mihalcea et al., (2008) empirical method is not transferable even to the same glacier under very similar meteorological conditions.
Although the total supraglacial debris load (accounting for void space) could not be calculated for the Mihalcea et al., (2008) debris thickness map from 2005 (as the exact debris thickness estimate values they calculated were not available), the supraglacial load for the Mihalcea et al., (2008) method when applied to the 2004 image could be obtained. This value was then compared to supraglacial debris load estimates calculated using all the previous approaches.
Table 5.18: Total supraglacial debris load, adjusted for assumed void space (average glacier value of 38% applied)

<table>
<thead>
<tr>
<th>Year</th>
<th>Method</th>
<th>Total supraglacial debris load (adjusted for assumed void space) m³</th>
</tr>
</thead>
<tbody>
<tr>
<td>2005</td>
<td>Empirical approach - Linear equation</td>
<td>35.39</td>
</tr>
<tr>
<td>2005</td>
<td>Empirical approach - Quadratic equation</td>
<td>37.25</td>
</tr>
<tr>
<td>2005</td>
<td>Energy balance model - (z_o) 0.002 m</td>
<td>40.69</td>
</tr>
<tr>
<td>2005</td>
<td>Energy balance model - (z_o) 0.003 m</td>
<td>51.26</td>
</tr>
<tr>
<td>2005</td>
<td>Energy balance model – (SWR) calculated using each pixel slope/aspect value, (z_o) 0.001 m</td>
<td>53.52</td>
</tr>
<tr>
<td>2005</td>
<td>Energy balance model - (SWR) calculated using average slope/aspect value, (z_o) 0.001 m</td>
<td>67.62</td>
</tr>
<tr>
<td>2004</td>
<td>Energy balance model - (z_o) 0.002 m</td>
<td>37.07</td>
</tr>
<tr>
<td>2004</td>
<td>Energy balance model - (z_o) 0.003 m</td>
<td>45.14</td>
</tr>
</tbody>
</table>

Table 5.18 highlights that although the Mihalcea et al., (2008) method applied to the 2004 ASTER image did not perform well visually, the total supraglacial debris load calculated was much greater than values calculated by both empirical approaches (linear and quadratic) and the simplified energy balance model. The model which performed best was the energy balance model with inclusion of \(SWR\downarrow\), which was calculated using an average slope and aspect value. This also highlights that results obtained using the energy balance approach can be as reliable as those obtained using highly tuned empirical approaches, which are restricted in their application due to the requirement of extensive fieldwork.

However, due to the low supraglacial debris load values obtained by the ‘simplified’ energy balance model, it is clear that for more reliable debris thickness estimates, the application of a more physically based model is required, which includes variable \(SWR\downarrow\) and accounts for the variability of other parameters between pixels, such as the \(z_o\) value. Therefore, applicability of the model should be increased when issues of \(SHF\) calculation and \(z_o\) value applied are investigated further.
5.12. Summary

This chapter has tested and developed a method to estimate debris thickness using thermal band imagery using an ASTER AST08 image from 01/08/05. The method is based on solving the surface energy balance at each pixel using measured incoming radiation fluxes and wind speed on the glacier, with distributed temperature from the satellite image to find the unknown value of debris thickness. As it is based on a physical solution to the surface energy balance, it has the potential to be transferable both to other glaciers and to other dates on the same glacier. This is a distinct advantage over a previous empirically-based approach (Mihalcea et al., 2008), which is not transferable and requires extensive in situ measurements at the study site to calibrate the surface temperature-debris thickness relationship at the moment of the satellite overpass.

The pattern of debris thickness distribution across Miage Glacier derived from the energy balance model was similar to that of the empirical method. Because the empirical model was calibrated with a large number of in situ debris thickness measurements, this result provides validation for the energy balance approach. Furthermore, the transferability of the energy balance approach (and similarly the non-transferability of empirical approaches) was highlighted by the generation of a very similar debris thickness map using a second thermal satellite image from a different year (2004), while the empirical method generated an unrealistic debris thickness map, with some implausible (negative) values using the 2004 image input. However, further testing is required.

Validation of either approach using field thickness measurements proved difficult due to the enormous challenge of generating a representative sample of debris thickness in the large area (90 x 90 m) of an ASTER pixel. The availability of another debris thickness map generated empirically from extensive field data (Mihalcea et al., 2008) provided an alternative and more
reliable method of validation in this study. However, alternative data may not always be available.

Despite this successful application, however, a number of key assumptions and simplifications had to be made in the energy balance model. The first assumption was that of a flat glacier surface, meaning that the same $LWR_{↑}$ and $SWR_{↓}$ values were applied to every pixel from the value recorded at the weather station. Therefore, to test for the effect of this assumption the model was developed to become more physically-based, using each pixel’s slope and aspect derived from a DEM to calculate distributed values of $SWR_{↓}$ and $LWR_{↑}$. However, the results of this increased physical representativeness were mixed, with some unrealistic values of debris thickness generated. A possible explanation for this result is that with increasing physical representativeness, the greater the need for input data to be precise and accurate. This may not be the case with the DEM used, which is much older (1991) than the satellite images (2004, 2005), leading to the possibility that downwasting and movement of debris has significantly altered the pattern of surface slopes and aspects in the intervening years, leading to errors in the estimation of radiative fluxes.

The second simplifying assumption in the model was that air temperature remained constant across the glacier. But, calculations at sites located away from the LWS, generated unrealistic debris thickness values due to problems in estimating the sensible heat flux, which was most likely due to the wrong air temperature value being used. On a debris-covered glacier it is much more likely that 2 m air temperature varies as a function of surface temperature, increasing or decreasing in thicker or thinner areas of debris.

To address this problem, a regression approach utilising the relationship between air temperature and surface temperature, based on morning measurements at the AWS under clear
sky conditions, was applied. The calculation of air temperature for each pixel using a regression equation was successful at most pixel locations. However, it became clear that at some pixel locations the $SHF$ values were still too large. It is likely that at these locations, air temperature is underestimated, leading to a steep temperature gradient with surface temperature, which in turn affects the calculated $SHF$ values. Improving the understanding of the spatial pattern of 2 m air temperature and its relationship to debris temperature and thickness is an important area for future work.

Overall, this chapter has shown the development of a method that could aid in the monitoring of debris thickness and extent changes over time, which are vital components of global glacier monitoring strategies. Ultimately, debris thickness needs to be mapped across entire mountain ranges to provide input fields for models of glacier mass balance, to estimate melt rates under future climate scenarios. Therefore, to ensure its transferability and potential to monitor debris thickness and extent changes on debris-cover glaciers, the model needs to be applied on a different glacier, along with investigation into the issues identified while the model was developed such as the calculation of air temperature at different pixels. Once these issues have been addressed and the model successfully applied at another location, the potential of the model for application into global glacier monitoring programs such as GLIMS for the rapid monitoring of debris thickness changes can be demonstrated.
CHAPTER 6: CHANGES IN DEBRIS COVER EXTENT OVER TIME

Chapter aim: To use TERRA ASTER and Landsat data (MSS, TM ETM+) to map changes in debris cover extent over time, using both manual and semi-automatic approaches.

6.1. Introduction

A number of attempts have been made previously to identify debris-covered glaciers and their margins (e.g. Taschner and Ranzi, 2002; Ranzi et al., 2004; Bolch and Kamp, 2005). Such studies have contributed to the GLIMS project, which aims to (automatically) monitor glacier extent changes over time. However, the monitoring of changes in debris cover extent over time has previously been limited, with studies utilising a variety of different methods including the visual analysis of multispectral images (Bishop et al., 1995), using supervised classification (Shukla et al., 2009), and the geomorphometric analysis of digital elevation models (DEM) (Bishop et al., 2000; 2001), none of which combine the potential of both multispectral images and DEM approaches for rapid (semi-automatic) debris extent monitoring (Paul et al., 2004).

Supraglacial debris exhibit the same spectral characteristics as the surrounding terrain. Therefore, automated mapping of the debris-covered area of a glacier from multispectral satellite imagery is not straightforward (Paul et al., 2004; Shukla et al., 2009). To resolve this many previous studies have applied a manual approach through on-screen digitising of debris-covered areas (Stokes et al., 2007). Although results can be accurate it is very time consuming, especially for a large glacier, or if a large number of glaciers are being studied (Paul et al., 2004; Bolch and Kamp, 2005, Bhambri and Bolch, 2009; Bolch et al., 2010). To address this, recent studies have developed methods which are semi-automatic, with some combining multispectral satellite imagery through band ratios (Bolch and Kamp, 2005), and
others which have developed band ratio approaches further to include other ancillary data such as DEMs (Paul et al., 2004).

In this study, the extent of debris cover and its variation since 1975 on Miage Glacier was identified using both a manual and semi-automatic approach to determine the benefits and disadvantages of each. The manual approach utilised on-screen digitising, while the semi-automatic approach was based on that described by Paul et al., (2004). This semi-automatic method was selected as other semi-automatic methods by Taschner and Ranzi (2002), and Ranzi et al., (2004) were developed solely to identify the boundaries of debris-covered glaciers and, therefore, would not be suitable for identifying the actual extent of the debris cover.

The study of debris cover variations is important, as an increase in extent and thickness will modify the relationship between temperature and ablation, with an increase in the extent of thick debris covers resulting in reduced melting even if the temperature is increased (Diolaiuti, et al., 2003). However, the opposite will be encountered and ablation enhanced, if an increase in thin and patchy debris occurs. Therefore, debris-covered glaciers respond differently to climate changes compared to ‘clean’ glaciers, experiencing involving widespread thinning rather than retreat. Debris extent variations are closely linked to climate and the mass balance of the glacier. During periods of negative mass balance the amount of debris cover on a glacier will increase, through the melt out of existing englacial sediments (Diolauti et al., 2009) and reduction in glacier transport rates (Kirkbride, in press). However, during periods of positive mass balance debris covers will decrease in extent as glacier transport rates will be sufficient to transport the debris down glacier, and melt out of englacial sediments will not occur.

Consequently, a debris layer has a significant influence on both mass balance and the discharge characteristics of emerging meltwater streams (Mattson, 2000; Shukla et al., 2009).
The changing discharges that result from debris-covered glaciers must be monitored to predict the long-term availability of water resources, due to the dependence of many upon melt water for irrigation (Kayastha et al., 2000).

The monitoring of debris extent variations, therefore, links to surface elevation changes, because the spread of a debris cover on a glacier can have either a positive or negative impact upon surface ablation, or its spatial distribution depending upon its thickness. Consequently, this also links into methods for obtaining debris thickness estimations, and the requirement of monitoring programs to combine debris extent and debris thickness monitoring, which may provide explanation for patterns of surface elevation change on a glacier, both spatially and over time.

### 6.2. Remotely sensed data

Both Landsat (MSS, TM, and ETM+) and TERRA ASTER data were used in the analysis, as the longer archive of Landsat MSS, TM, and ETM+ provided images pre 2000.

#### 6.2.1. Landsat imagery

Three cloud free images were used spanning 24 years so that the development of the debris cover over a reasonably large timescale could be investigated: 13/07/1975 (Landsat MSS, Landsat 1), 10/09/90 (Landsat TM, Landsat 5), 25/07/99 (Landsat ETM+, Landsat 7). However, because the Landsat MSS sensor has only four bands in the visible/near infra red and no thermal band it could not be used in the semi-automatic method. The choice of images was restricted due to the presence of snow and/or cloud on a number of images (with snow frequently completely covering the glacier between October-May), which limited the number of images suitable for use. Also images were obtained freely from an orthorectified Landsat data set which had a restricted number of free images available (Landsat 2008).
6.2.2. ASTER imagery

Four ASTER images were obtained from 02/07/00, 14/08/04, 01/08/05, and 26/06/06, potentially providing more recent information on debris extent. However, some of these images (2005, and 2006) had significant amounts of snow and cloud present which limited debris extent analysis.

6.3. Methodology

The first part of this chapter will focus upon the methodology used for both manual digitising and the semi-automatic method.

6.3.1. Manual delineation

Onscreen digitisation was used to manually delineate the debris-covered areas on the glacier surface. Areas covered by debris were highlighted as an area of interest (AOI) in the ERDAS IMAGINE image processing software by drawing a digitised line around the area (Figure 6.1 b). Areas of debris were visually identified due to their darker appearance on the images compared to the lighter/whiter bare ice areas (Figure 6.1 a). The minimum mapping unit was a group of pixels, rather than at a single pixel scale which was difficult to discriminate.

The precision of this method was tested using a method previously applied by Stokes et al., (2007) to a study of debris extent changes in the Caucasus, Russia. The precision was calculated after the debris outline was digitised 10 times and the maximum distance between any two digitised lines recorded. Delineation should be performed by one person so that it remains consistent (Bolch et al., 2008), and for further accuracy identification some fieldwork is required (Stokes et al., 2007).
Figure 6.1: a) Identification of debris areas showing clear difference between dark debris-covered areas, blue-grey (ice) and white (snow) non-debris covered areas,  b) manually digitised areas of debris cover, shown in green (ASTER false colour composite image R:1, G:2, B:3N, 14/08/04).

### 6.3.2. Semi-automatic method

The manual delineation of multispectral images can be difficult, subjective, and time consuming, with the resulting accuracy limited due to the problems of both mixed pixels and presence of fine surface dust or sporadic thin debris (Paul et al., 2004; Bolch and Kamp, 2005, Bolch et al., 2010). Therefore, the semi-automatic approach of Paul et al., (2004) was applied. This combines multispectral data with elevation data derived from a DEM, and was successfully applied on the Oberaltschgeltscher, in the Swiss Alps. In this method a number of different steps were completed, each excluding areas on the image which were not of interest (not debris-covered). This results in a final image showing just the debris-covered areas of a glacier. Variation in debris extent can be identified from applying this procedure to multi-date images.

1) The first step attempted to minimise the impacts related to the variation in topography, such as anisotropic reflectance, by producing a band ratio image using TM4/TM5 or ASTER 3/4 (Mather 2004, Paul et.al. 2004).
2) This ratio image was then divided into two classes of ‘glacier’ (black) and ‘other’ (white) areas classified using a threshold value (of 2) published by Paul *et al.*, (2004). The threshold value was verified by visual inspection, indicating that when the threshold is reduced the greater the number of partly debris-covered pixels were included, but at the expense of more noise elsewhere in the image (Paul *et al.*, 2004).

3) An Intensity Hue Saturation image (IHS) was produced using bands TM3, TM4, and TM5 or ASTER 2, 3, 4. This enhanced the three channels in relation to their colour contrast (Mather, 2004) and separated the debris from other features in the image.

4) The Hue component of IHS image (channel 2) was used to map vegetation (black) and vegetation free areas (white) using a threshold value identified by Paul *et al.*, (2004) of 126, which was visually validated.

5) Using a DEM, all slopes greater than 24° were excluded.

6) Each of the maps were overlain, masking those areas which were not debris-covered so that only debris-covered ice areas were shown in white.

### 6.4. Method testing and development

Both the manual and semi-automatic approaches were tested for their applicability on the Miage Glacier and any problems with this application identified and solutions presented.

#### 6.4.1. Manual digitising

Debris extent was manually extracted for each of the three Landsat images (MSS, TM, and ETM+) and four ASTER images so that changes between 1975-1999 could be identified. However, the presence of snow and clouds on the 2005 and 2006 ASTER images, and coarse spatial resolution of the Landsat MSS image restricted the successful delineation of debris extent further up the glacier in these images, and in particular the critical position of the upper limit of continuous debris cover (the area where greatest changes are likely to have occurred).
6.4.1.1. Sporadic debris problems and its treatment

A number of problems emerged during the onscreen digitising of debris cover. The first occurred on the areas of the glacier (especially the upper reaches) where the debris is very patchy and sporadic, making it difficult to draw around without including patches of bare ice near by (Figure 6.1a). Difficulties also arose due to the presence of mixed pixels of bare ice and sporadic debris, with the mixture of bare ice/debris occurring at a sub-pixel scale making discrimination impossible.

Figure 6.2: Field photographs illustrating debris-free areas in the upper reaches of the glacier and also the problem of snow cover at the upper limit of debris cover, a) Tributary Glacier de Mont Blanc, July 2007, b) upper limit of debris cover on the Miage Glacier, June 2005, c) upper limit of debris cover on the Miage Glacier, June 2006, d) upper limit of debris cover on the Miage Glacier, June 2007.
Another problem that became apparent when looking at the satellite images was the presence of fine surface dust. This was apparent on the upper glacier (and its tributaries), which appeared darker than the bare ice areas, implying that a debris layer was present (Figure 6.1 a). After checking these areas with recent field photographs (Figure 6.2 a to d), it was apparent these extensive areas were not covered with sporadic debris but consisted of bare ice covered with fine surface dust. The presence of this fine surface dust on bare ice creates a lower albedo, and makes these locations appear darker on the satellite image (Brock et al., 2000).

This highlights a key problem with the manual delineation of debris cover extent (especially on images acquired late in the ablation season when dust on the bare ice is at its greatest). Because, without fieldwork, areas of clean ice could have been mis-classified as patchy debris cover, highlighting the subjectivity of this method. In turn, this meant that all debris had to be classified as one class, as the sporadic debris could not be discriminated from the continuous debris cover on the Landsat TM and ASTER imagery due to the spatial resolution being too coarse. The ability to distinguish the sporadic debris would be useful as it has implications upon mass balance calculations, due to melt being enhanced in locations where the debris is thin and patchy. To enable this, imagery with a finer spatial resolution would need to be tested to see whether sporadic debris could be detected.

6.4.2. Semi automatic method

The semi-automatic method of Paul et al., (2004) was initially applied to a Landsat TM and (10/09/90, Figure 6.3 a-f) and ASTER (14/08/04, Figure 6.4 a-f) image to ensure its application would produce reliable results.
Figure 6.3: Processing steps completed for debris extent identification on Landsat TM 10/09/90 image, a) TM4/TM5 band ratio, with glacier areas white, b) TM4/TM5 band ratio image with threshold of 2 applied, glaciers white, c) IHS image of TM band 3,4,and 5, d) IHS channel 2 (hue image) with threshold of 126 applied, therefore, all vegetation removed, e) slope image showing areas <24° (white), f) debris extent (white).
Figure 6.4: Processing steps completed for debris extent identification on ASTER 14/08/04 image, a) VNIR3/SWIR4 band ratio, with glacier areas white, b) VNIR3/SWIR4 band ratio image with threshold of 2 applied, glaciers white, c) IHS image of TM band 3,4, and 5, d) IHS channel 2 (hue image) with threshold of 126 applied, therefore, all vegetation removed, e) slope image showing areas <24° (white), f) debris extent (white)
6.4.2.1. Method Problems

Once the semi-automatic method had been applied to both ASTER and Landsat TM images, a number of problems were evident. First, on the debris-covered tongue a number of linear features running down glacier are shown as debris-free ice (shown as black, Figure 6.3 and 6.4 e and f). These areas are in fact debris-covered, and are the lateral slopes of the medial moraines which are present along the glacier tongue. These features are excluded due to their high slope angle, which at some points on the glacier are greater than 24° and up to 45° in some locations (Figure 6.5 and 6.6). Therefore, different slope angles were applied to see if these features could be included by increasing the slope angles, and what the resulting impact would be upon other areas.

Figure 6.5: Photograph showing steepness of medial moraines, Miage Glacier.
Figure 6.6: 1991 Orthophoto DEM slope map showing high lateral slope angles of the medial moraines.

By increasing the slope angle, the moraines were included in the final debris extent image (6.7 a to f) however, the amount of white ‘speckle’ (debris-covered areas) on the steep mountainous terrain and valley bottom increased. The detail of the glacier snout extent was completely lost when slope angles up to 45° were retained, but this did successfully identify all medial moraine areas (Figure 6.7 e).
Figure 6.7: Application of different slope angles to the orthophoto DEM (step 5 of semi-automatic classification), a) orthophoto with slopes >24° removed, b) debris extent map with slopes >24° excluded applied to ASTER 2004, c) orthophoto with slopes >30° removed, d) debris extent map with slopes >30° removed applied to ASTER 2004, e) orthophoto with slopes >45° removed, f) debris extent map with slopes >45° removed applied to ASTER 2004.

Because the reason for these dark bands on the glacier was known, and that they are known to be debris-covered, it was decided to leave the slope angle at 24°. However, this clearly shows a problem of generalization between different glaciers. A solution would be to manually
include these areas as debris-covered following inspection of the final debris extent images. This was also identified by Paul et al., (2004) and identified as the ‘correction of remaining artefacts’ and was applied as a final processing step. The best band combinations for this identification are TM5, TM4, TM3 (ASTER 4, 3, 2) along with an overlay of automatically derived glacier areas (Paul et al., 2004). This initial sensitivity to slope angle was completed on a high spatial resolution orthophoto DEM, but as this did not cover the full extent of the upper reaches of the glacier debris extent could not be fully identified in the upper reaches. Therefore, an ASTER DEM (02/07/00) which covered the entire glacier was also applied and its ability to estimate debris extent tested.

When applied to the coarser spatial resolution ASTER DEM (30 m), more debris-covered areas in the upper part of the glacier were omitted, as were the medial moraines whose slope angle was steeper than the critical angle of 24 ° (Figure 6.8 a to f). Increasing this angle resulted in more debris-covered areas being retained, but with the effect of including more of the surrounding (and unwanted) regions (Figure 6.8 b, d, and f). The coarser spatial resolution of the ASTER DEM explains the reduction in visible detail in the debris extent maps compared to results using the 10 m orthophoto DEM, a clear indication that the accuracy and detail of output of the semi-automatic method is heavily dependent on having a high resolution DEM of at least 25 m (as used by Paul et al., 2004).

Although the debris extent maps using the ASTER DEM are less detailed they do give a good synoptic view of the whole glacier. They could also be more useful on the larger debris-covered glaciers in the Himalayas or Andes (Paul et al., 2004; Bhambri and Bolch, 2009).
Figure 6.8: ASTER DEM (02/07/00)  a) slope image with all slopes >24° removed, b) debris extent image with ASTER DEM slope image applied (slopes >24° removed) to semi-automatic method, c) slope image with all slopes >30° removed, d) debris extent image with ASTER DEM slope image applied (slopes >30° removed) to semi-automatic method, e) slope image with all slopes >50° removed, f) debris extent image with ASTER DEM slope image applied (slopes >50° removed) to semi-automatic method.
6.5. Debris extent results

Both methods were applied to all of the available images (both ASTER and Landsat), although due to the lack of middle infrared wavebands the semi-automatic method was not applied to the Landsat MSS image. The semi-automatic method was applied using a slope threshold of 24° so that the inclusion of surrounding terrain would be minimal. An orthophoto DEM from 1991 was used despite it excluding upper parts of the glacier, due to accuracy and spatial resolution problems with the more recent ASTER 2000 DEM. However, due to the problems of snow cover, cloud and spatial resolution (Figure 6.9 a and b), only the 1990 and 2004 (Figure 6.10) images could be used to estimate changes in debris extent with any degree of reliability.

![Figure 6.9](image)

**Figure 6.9:** a) 1975 Landsat MSS satellite image highlighting problems of both coarse spatial resolution and snow, b) 2006 ASTER image showing issues of snow and cloud cover which obscure the glacier.

<table>
<thead>
<tr>
<th>Year</th>
<th>Debris area – manual method (km²)</th>
<th>Debris area – semi-automatic method (km²)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1990</td>
<td>4.15</td>
<td>3.88</td>
</tr>
<tr>
<td>2004</td>
<td>4.25</td>
<td>4.06</td>
</tr>
<tr>
<td>% Change</td>
<td>+ 2.41</td>
<td>+ 4.64</td>
</tr>
</tbody>
</table>
Results from the debris extent images were compared to field photographs from years corresponding to the more recent ASTER images. With the 14/08/04 ASTER image outputs compared to 2005, 2006, and 2007 field photographs.

### 6.5.1. Manual digitising: Precision analysis

In an approach similar to that of Stokes et al., (2007) the precision error of the manual method of on screen digitising was identified by digitising the debris covered area 10 times (Figure 6.10).
Next the maximum, minimum, and average precision error was identified by measuring the distance between each of these digitised lines at a number of different points (25 in total) (Table 6.1).

**Table 6.1:** Precision uncertainty in manual digitising, with the maximum and minimum errors being the largest and smallest distance between any of the digitised lines at random sample locations (25 locations in total).

<table>
<thead>
<tr>
<th>Year</th>
<th>Maximum (m)</th>
<th>Minimum (m)</th>
<th>Average (m)</th>
<th>Standard deviation (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1990</td>
<td>160</td>
<td>24</td>
<td>46</td>
<td>29</td>
</tr>
<tr>
<td>2004</td>
<td>107</td>
<td>21</td>
<td>49</td>
<td>26</td>
</tr>
</tbody>
</table>

The greatest error occurred at the boundary between the debris and bare ice in the upper reaches, due to the greater occurrence of mixed pixels and patchy debris cover (Figure 6.12). Also, results from the precision testing (Table 6.1, Figure 6.11) show that the precision errors for 1990 and 2004 are very similar. A reason for this may be the similar spatial resolutions of the two sensors used which meant areas (such as the glacier boundary and debris/ice interface) were difficult to digitise in both images, due to the occurrence of mixed pixels at the same spatial scale.
Some of these precision errors are greater than those found by Stokes et al., (2007) on a similar study in the Caucasus Mountains, Russia, who found maximum precision errors of +/- 25 m. The largest errors tended to be found in the upper reaches in the transition zone from debris-covered to bare ice. To address these errors, a finer spatial scale could be used to ensure the detail in small debris patches in the upper reaches are digitised precisely. However, the problem of mixed pixels will still make this difficult even at a finer (single pixel) scale. However, despite these errors, the method of manual delineation does provide a means for determining the extent of debris extent on a glacier which can be verified with fieldwork in a much quicker, easier and cheaper way than a method based upon fieldwork alone, which can also experience the same problems of the identification of the debris margins (Stokes et al., 2007).

6.6. Analysis of results

6.6.1. Glacier Snout

Both the manual and semi-automatic methods identified that the snout was completely debris-covered, and photographs verify these results (Figure 6.13). Only limited ice exposure was
identified where isolated ice cliffs are present, which also concurs with field experience of the areas.

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**Figure 6.13:** *Lower reaches a) 2004 debris extent from Paul et al., (2004) method, b) 2004 debris cover manually delineated (green debris cover), c) completely covered glacier snout (looking down glacier) June 2006, d) completely covered glacier snout (looking up valley towards the glacier) August 2007.*

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### 6.6.2. Middle reaches

Photographic and field evidence suggest that the middle reaches of the Miage Glacier are completely debris-covered, with bare ice exposed only on isolated ice cliff faces. This bare ice was not picked up by the two methods due to the spatial resolution of the ASTER image at 15 m, which is much coarser than smaller scale features such as exposed ice cliffs. Also, their orientation (with the bare ice face at a very steep vertical angle) means they will not be viewed by the sensor.
Figure 6.14: Middle reaches a) 2004 Debris cover of middle reaches using Paul et al., (2004) method, b) 2004 debris cover manually delineated, c) completely covered medial moraine June 2006 (1), d) completely covered areas away from the medial moraine June 2006 (3), e) bare ice on exposed cliff faces August 2007 (2).

6.6.3. Upper reaches

Figure 6.14 (c and f) provides evidence to verify that debris is patchy in the upper areas of the glacier, with sporadic coverage where tributary glaciers feed into the Miage Glacier. Most debris is concentrated into two dominant medial moraines which continue down glacier to the snout. This was most clearly identified by the semi-automatic method (Figure 6.15 a).
Figure 6.15: a) 2004 debris cover from upper reaches using Paul et al., (2004) method b) debris cover manually delineated (green = debris, yellow = bare ice), c) bare rock between tributary glaciers – glacier de Mont Blanc 2005 (1), d)-e) 1990 and 2004 Satellite images showing area of bare rock on the slopes of the glacier de Mont Blanc appears to look debris-covered (area circled in red) , f) medial moraines – debris covered August 2007 (2).

One area of misclassification was evident on the manual method of digitising where an area of bare rock (located at the glacier de Mont Blanc) has been incorrectly classified as debris-covered due to the similar appearance of debris and rock material on the image. The semi-
automatic method does not misclassify this area due to its very steep slope angle. Therefore, on the debris output image it was clearly identified as non-debris covered (Figure 6.15 a).

6.7. Discussion of results

It was noted that there was a small increase in debris extent between 1990 and 2004 using the manual method (0.10 km² over 14 years) (Table 6.2) but its significance is questionable given the errors associated with this method discussed previously. However, the semi-automatic method also shows a slight increase (0.18 km² over 14 years). The semi-automatic method has shown a slightly larger increase in debris cover extent, and may be due to precision errors in the manual method of digitizing, which may miss some debris covered pixels due to human error in visually identifying the debris cover.

Figure 6.16: Debris cover development of the Miage Glacier between 1770-1940, Key: 1 – Clean ice, 2 – discontinuous debris cover, 3 – continuous debris cover, 4 – medial moraine, 5 – local rock-avalanche deposit (Deline, 2005).
The fact that there is only evidence of a small increase in debris cover suggests that most of the increase in debris cover on the Miage Glacier tongue occurred prior to the 1990s. This is supported by Deline (2005), who suggests that the Miage Glacier debris cover was similar to its present extent in the 1940s (figure 6.16).

Despite not finding evidence to suggest a large increase in debris cover has occurred on the Miage glacier since 1990. The semi-automatic method of Paul et al., (2004) has been successfully applied to the Miage Glacier, and to both Landsat TM and ASTER data, with the main debris-covered areas clearly visible on the debris extent maps. The method of manual digitising has also been applied with some successes, with error analysis showing positional uncertainty in general averaging around 40 m. Despite this being a highly subjective method, it can be useful in determining the general pattern of debris cover changes over time (Stokes et al., 2007).

6.8. Summary

This chapter has focussed on the identification of debris extent variations on the Miage Glacier. To identify the changes in debris extent over time, two different methods were utilised. First, a manual method of digitising which was time consuming and labour intensive, and second, a semi-automatic approach based on that of Paul et al., (2004). Results identified that only a small increase in debris cover extent occurred during the study period (1990-2004), highlighting that the rapid increase in debris cover occurred prior to this.

One problem that became evident in both manual digitising and the semi-automatic method was the presence of cloud and/or snow cover within the satellite images, which hampered full investigation as it obscured the debris cover and ice beneath. This highlights the requirement for high quality cloud and snow free images from the late season to minimise snow and cloud
cover issues. However, the problems of debris cover identification in the upper reaches (where debris is patchy) will be increased at this time, as the glacier ice will be dark due to the incorporation of a season of dust fall. This highlights that in some cases, the acquisition of a number of images without cloud and/or snow cover is difficult due to the climatic conditions of the area of study.

Also, due to the spatial resolution of the ASTER and Landsat imagery, thin and patchy debris could not be identified with either method. These areas are of key interest in mass balance studies due to the enhanced ablation under these debris layers, highlighting the requirement for finer spatial resolution imagery at these locations (the debris/ice margin). Another issue was the repeat time-scale of the sensor (ASTER 48 days, Landsat MS 18 days, TM 16 days, ETM+ 16 days), which limits the number of images per year (although some sensors can be tasked to increase the revisit time) that can be used.

The semi-automated method of Paul et al., (2004) gains strong support from this study with successful application to the Miage Glacier and both Landsat (TM and ETM+) and ASTER data. However, results need manual checking due to the misclassification of steep moraines. This method is also better at discriminating clean ice from debris-covered ice when compared to the approach of manual delineation, which can incorrectly identify bare ice as debris if it appears slightly darker on the image (due to the present of surface dust). Also, more patchy debris-covered areas are discriminated better using the semi-automatic method and is difficult to determine manually.
CHAPTER 7: MONITORING SURFACE ELEVATION CHANGES AND VELOCITY AS A MEANS OF ASSESSING THE MASS BALANCE STATUS OF DEBRIS-COVERED GLACIERS

**Chapter aim:** To use ASTER DEMs and visible imagery to monitor both changes in surface elevation over time, and estimate surface velocities using feature tracking methods.

### 7.1. Introduction

Surface elevation change is controlled by surface ablation and ice flux, therefore, any changes in surface elevation highlight that a change has occurred in relation to the ice flux into the glacier and/or surface ablation levels. Both surface ablation rates and ice flux are controlled by climate, therefore, any changes in surface elevation can identify a change in climate. Consequently, the accurate determination of glacier surface elevational change over time provides a means of identifying the impacts of climate change upon glacier mass balance (Salerno *et al.*, 2008), assessment of glacial fresh water resources they represent, and the potential contribution of glaciers to sea level rise (Berthier *et al.*, 2004).

At present, only a small number of glaciers are being monitored for mass balance changes by direct methods, with even fewer long term monitoring programs due to access problems and economic limitations (Berthier *et al.*, 2004; Haeberli *et al.*, 2007). There has previously been a bias towards measuring large glaciers and their snout length variations, with changes in elevation being neglected especially on small mountain glaciers (Paul *et al.*, 2007).

Although changes in glacier thickness cannot be measured directly using satellite images, the analysis of a time series of images provides evidence of volume changes over time by DEM
subtraction (Paul et al., 2007). In turn, this provides an indirect measurement of mass balance, because change in surface elevation over time can be an indicator of mass balance changes (Etzelmuller, 2000; Racoviteanu et al., 2007). The ability to measure thickness changes using remote sensing enables more extensive glacier monitoring than previously possible via methods such as aerial photographs, and direct glaciological measurements due to the large spatial coverage of remotely sensed images.

Consequently, this will enable the rapid and repeated analysis of the mass balance variations of an increased number of glaciers across the globe (Berthier et al., 2006). In relation to debris-covered glaciers, elevation change monitoring also provides a better method of monitoring the state of the glacier and its response to changes in climate. Unlike clean glaciers, the frontal recession of a debris-covered glacier is a poor indicator of its negative mass balance, as downwasting is the most common source of losses (Luckman et al., 2007). Also, future monitoring programs could combine surface elevation changes analysis with the debris thickness model identified in chapter 5, to identify the role of both debris thickness and extent in the spatial distribution of elevation changes.

The ice flow of a glacier is controlled by inputs (accumulation) outputs (ablation), slope, temperature, and ice transport of the glacier system (Kaab, 2005). The monitoring of ice velocity provides a means of understanding glacier dynamics (Vadon and Berthier, 2004) and, as a result, the balance status or ‘health’ of a glacier can be identified. The monitoring of a glaciers velocity is, therefore, essential because a decrease in velocity due to a reduction in the input of ice into the glacier in the accumulation zone will result in glacial stagnation, and the warming trend of today’s climate will increase the stagnation (stationary downwasting) of many debris-covered glaciers (Kaab, 2005; Luckman et al., 2007). Remote sensing techniques using both optical and microwave Synthetic Aperture Radar (SAR) have frequently
been applied successfully to estimating ice velocities both on clean (Kaab, 2005; Kaab et al.,
2005; Bevan et al., 2005) and debris-covered glaciers (Kaab, 2002; Vadon and Berthier, 2004;
Berthier et al., 2005; Luckman et al., 2007).

This chapter, therefore, has two main aims. First, to apply TERRA ASTER satellite data to
identify surface elevation changes and investigate its potential use for mass balance
monitoring. Second, the potential of ASTER data (both multispectral and DEM data) will be
investigated in relation to estimating surface velocity measurements.

### 7.2. Elevation data

To identify elevational changes over time, the geodetic method was applied, which compares
DEM s from different years, with the thickness change at each pixel identified by subtracting
one DEM from another (Etzelmuller, 2000). In this study ASTER DEM (AST14-relative) data
was used (section 4.5). These DEMs were relative, meaning data could not be compared with
direct field measurements of elevation, however, this limitation was a not a problem for
assessing vertical elevation change between two ASTER Relative DEMs.

#### 7.2.1. Error analysis

Before the ASTER DEM was used to identify vertical surface changes over time, the level of
geographical and vertical accuracy was tested.

#### 7.2.1.1. Geographical accuracy

Geographical accuracy needed testing to confirm that each of the ASTER DEMs were
registered to each other, ensuring that change between DEMs was due to elevation change, not
g elo locational errors. To test this, the location of a known point was visually compared on
each of the ASTER DEM orthorectified visible channels (as the DEMs spatial resolution is too
coarse to identify features on or around the glacier) to ensure it was in the same place. This known point was the location of a GPS base station. Additionally, the GPS derived locations of the 25 ablation stakes (2005 fieldwork) on the glacier were visually compared in each DEM, to ensure they were in the same place.

The 2000, 2004, and 2005 DEM showed no variation in the location of the base station or the 2005 ablation stakes. However, the 2006 visible orthorectified channels (and resulting DEM) showed slight variation in position to the other datasets (Figure 7.1 b). To resolve this, the 2006 orthorectified channels and DEM were manually georeferenced to one of the other images known to be correctly georeferenced (2000 visible channel – orthorectified to the 2000 DEM). This was completed using image processing software (ENVI) and a 3rd degree polynomial transformation, with 25 ground control points distributed across the image to ensure correct georeferencing throughout the image. A 3rd degree polynomial was used due to the terrain in the study area, as a higher number of ground control points (and resulting polynomial) is required in areas of steep terrain (Mather, 2004). Once the 2006 visible channels and DEM were georeferenced with a Root Mean Square Error (RMSE) of 1.41 pixels, the locations of the 2005 stakes and GPS base station were again checked and found to be correct (Figure 7.1c).
Figure 7.1: Location of 2005 stakes and 2007 GPS base station on a) 2000 ASTER image, b) 2006 ASTER image showing geolocational error on 2006 image, c) 2006 (newly georeferenced) images showing removal of geolocation errors from the 2006 ASTER image
7.2.1.2. Vertical accuracy

To identify the level of vertical accuracy between the images, elevation values from each of the ASTER DEMs in areas of flat and steep terrain outside of the glaciated area (where no significant elevation changes should have occurred) were compared to assess the amount of difference (if any) which occurred between the DEMs (Figure 7.2). This approach was selected because the ASTER DEMs were relative so they could not be compared to GPS elevation points (absolute heights) taken in the field.

![Locations of sample points on both flat and steep terrain.](image)

Table 7.1 highlights significant problems with two of the DEMs in terms of their vertical accuracy, with both the 2005 and 2006 DEMs producing significant discrepancies compared to other DEMs both on flat and steep terrain. These differences are significantly greater than the values published by the ASTER processing team of 30 m (Lang and Welch, 1999). The 2000-2004 images produce a low average difference of -17 m and -30 m on flat and steep areas respectively, clearly within the published vertical range of the ASTER sensor of 30 m.
Despite these low average values, when the maximum elevation differences (both positive and negative) are analysed, the values on steep terrain are significantly larger than the 30 m value identified by the ASTER team.

**Table 7.1:** Average, maximum negative, and maximum positive differences experienced at 30 sample points between each DEM a) flat ground, b) steep terrain

<table>
<thead>
<tr>
<th></th>
<th>Average error (m)</th>
<th>Standard deviation</th>
<th>Max -/ive difference (m)</th>
<th>Max +/ive difference (m)</th>
<th>RMSE (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>a)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>2000-2004</td>
<td>-17</td>
<td>14</td>
<td>-44</td>
<td>18</td>
<td>22</td>
</tr>
<tr>
<td>2004-2005</td>
<td>567</td>
<td>387</td>
<td>-17</td>
<td>1301</td>
<td>683</td>
</tr>
<tr>
<td>2005-2006</td>
<td>1211</td>
<td>621</td>
<td>-4</td>
<td>2514</td>
<td>1356</td>
</tr>
<tr>
<td>b)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>2000-2004</td>
<td>-30</td>
<td>86</td>
<td>-432</td>
<td>290</td>
<td>89</td>
</tr>
<tr>
<td>2004-2005</td>
<td>129</td>
<td>350</td>
<td>-59</td>
<td>1502</td>
<td>368</td>
</tr>
<tr>
<td>2005-2006</td>
<td>555</td>
<td>490</td>
<td>-298</td>
<td>1555</td>
<td>740</td>
</tr>
</tbody>
</table>

The large differences in the 2000-2004 DEMs (although no way near as large as the 2005-2006 DEMs) can be explained by the steep and varied topography which generates significantly greater errors (Kaab, 2002; Bolch et al. 2008). Kaab (2002) also noted this problem when comparing an ASTER DEM to a reference DEM in Gruben, Swiss Alps, and found vertical differences of 60 m RMSE in complex steep and varied terrain, but lower differences of 18 m RMSE were identified in a section of more moderate terrain. Therefore, because the glacier tongue is relatively flat, snow free and unshaded, minimum errors are most likely to be experienced here.

Kaab (2002) also highlighted that the presence of additional factors such as snow patches, deep shadows, and steep rock walls in either moderate or complex terrain also has an impact upon the quality of the DEM generated. Areas containing all of these factors were identified as ‘worst case scenarios’ in DEM generation (Kaab, 2002). Bolch *et al.*, (2008) found much
greater errors than Kaab (2002) on steep slopes and where snow cover was present, with values up to 300 m, in the Khumbu Himal, Nepal. This highlights that the quality of the DEM elevation values varies significantly depending upon the topography of the study site, and whether snow and cloud is present. This lead Bolch et al., (2008) to conclude that accuracy values of 30 RMSE are acceptable in areas of steep terrain (with high relief differences), due to the problems generated by the terrain for the DEM generation.

Toutin (2008) looked at a number of different ASTER DEM studies and, in conclusion, found two different outcomes when using ASTER DEMs in mountainous regions. The first, where elevation ranges of more than 2000 m are found, and accuracies of +/- 15-30 m with a 68% confidence level can be achieved, however, post – processing is required to achieve this. Second, in difficult high mountain conditions where elevation ranges of 0-5000 m occur along with steep rock walls and snowfields where accuracies of +/- 60 m with 68% confidence level can be obtained. The second case scenario applies to the Miage Glacier study site, with snow, steep valley rock walls and high elevational ranges. The larger errors on the 2005 and 2006 DEMs result due to the presence of snow and cloud patches (Figure 7.4 a and b), which when combined with the steep topography, produce errors greater than those in the 2000-2004 images.
Figure 7.3: Location of sample points and difference values found at each site 2005-2006, showing increase in difference where snow, cloud, or shadow from the clouds is present, a) flat terrain, b) steep terrain.
7.2.1.3. Elevation differences related to slope and aspect

Kaab (2002) and Toutin (2008) noted that errors were concentrated on northern or north-western slopes greater than 35 °, due to them being hidden from the backward looking waveband (3B), as they lie in shadow. Also, the greater the slope, the higher the level of error generated, due to local topography effects and the bidirectional reflectance distribution function (BRDF) (Toutin, 2008). Racoviteanu et al., (2007), however, found that errors were not just concentrated onto northern slopes but also occurred on aspects between 0-180 °, meaning that the errors experienced cannot be solely the result of the backward looking channel. Therefore, to see whether a similar pattern was present the slope and aspect of each sample point used above (Figure 7.3) were extracted. Next, the elevation change errors were plotted against slope and aspect to see whether a trend was apparent.

It was identified that between 2000-2004, the greatest differences (>25 m) were found on slopes facing a northerly and a south to south easterly direction (0 and 150-180 °), and at slope angles of 30-40 ° (Figure 7.4 a and c). The range of aspects identified in the 2000-2004 DEMs correspond to findings Racoviteanu et al., (2007) who also found errors occurring on aspects ranging 0-180 °. The frequency of differences on the 2000-2004 DEMs were much lower (only 6 locations > 25 m) than the 2004-2005 difference results (18 locations > 25 m). Differences in the 2004-2005 and 2005-2006 DEMs were also much more widespread, and occurred between 0-200 ° in aspect (north to north north west) and on slopes ranging from 0 to 50 °. Similar differences were found by Racoviteanu et.al. (2007) who identified that the greatest errors during their study (~100 m) occurred on steep slopes around 60-77 °.
7.3. Extracting elevation changes

To extract changes in elevation on the Miage Glacier, the glacier area was identified by manually drawing around the glacier margins and creating an area of interest (AOI) in ERDAS IMAGINE. Once an AOI had been identified (Figure 7.5 a and b) it was extracted from all ASTER DEMs as an ASCII file. This created a text file containing the location and height of every pixel within the AOI. Once this was completed, the difference between the two DEMs could be obtained. This meant the changes in elevation between the two dates were identified for every ASTER pixel, with any thickening highlighted as positive and thinning as a negative (Berthier et al., 2004). Finally, the results were input into a GIS to identify the spatial pattern of elevation change.

Figure 7.5: Area of interest a) 2000 and 2004 areas, b) 2005, showing areas where data was excluded due to obvious distortions in the 3B band which resulted due to the presence of cloud and snow.

7.4. Surface elevation changes

Surface elevation changes between 2000-2004 and 2004-2005 were plotted to identify the distribution of elevation changes with elevation (Figure 7.6).
Looking at Figure 7.6 (a and b), significant errors were evident, with hundreds of metres of elevation change identified as occurring between 2000-2004 and 2004-2005. The 2000-2004 plot shows that the errors in elevation change increase (both positively and negatively) with an
increase in elevation, with positive over-estimates occurring above 2150 m, and negative over-estimates above 2400 m (located in the glacier upper reaches).

Therefore, all data above 2150 m were excluded and the elevation change results plotted again to see if a pattern were evident in the results (Figure 7.7a). This was repeated for 2004-2005,
where all data above 2100 m were excluded (Figure 7.7b), as over estimates occur above this point. The fact that problems with the elevation estimates occur above similar elevations in both image comparisons highlights that it is a problem with data above this point and was a result of the limitations of ASTER in a deep steep-sided valley.

The 2000-2004 data below 2150 m (Figure 7.7a) shows a general trend of greater thinning with an increase in elevation. This corresponds to the thinner debris cover noted up glacier, which unlike the thicker debris cover in the middle and lower reaches of the glacier (which insulates the ice beneath protecting it from melting), accelerates ice melt. However, the 2004-2005 data (up to 2150 m) (Figure 7.7b) does not show this trend, and shows an unrealistic trend of positive elevation change with an increase in elevation. A consequence of the large errors present within the 2005 DEM, which increase with an increase in elevation (due to the increase in steep and varied terrain with this elevation increase). Therefore, compared to the 2000-2004 data, this is unlikely to be an accurate representation of what is actually happening with regard to glacier surface changes.

Elevation changes were plotted in a GIS (Figure 7.8) to assess the spatial distribution of both positive and negative elevation changes and, if unexpected elevation change values were concentrated into certain areas. The 2000-2004 data were divided by 4 to give annual rates to enable comparison with the 2004-2005 results. Also, areas where snow, cloud, or terrain caused distortion on the images were removed.
Figure 7.8: Surface elevation changes a) 2000-2004 (annual rates), b) suitable glacier area 2004-2005, c) data up to 2150 m 2000-2004 (annual rates), d) data up to 2150 m 2004-2005 (all negative changes shown in red, all positive in green).
7.4.1. Discussion of results

When these findings are compared to previous studies on the Miage Glacier, some similarities in the general patterns identified occur, such as the thinning in the upper reaches. The pattern of thickening identified in the lower reaches between 2000-2004, is also noted by Smiraglia et al., (2000), and Diolaiuti et al., (2009). However, despite these similarities in the general patterns of surface elevation change, when the actual values of elevation change are compared, it is clear that the values of elevation change identified from the ASTER DEMs are unrealistic. In the lobe region for example many areas were identified of having an increase of greater than 4 m y$^{-1}$ between 2000-2004 ad 2004-2005, which is significantly higher than the 1 m y$^{-1}$ (40 m between 1975 and 1999) (Smiraglia et al., 2000) and 1.7 m y$^{-1}$ (20 m between 1991 and 2003) (Diolaiuti et al., 2009).

Therefore, due to the presence of significant elevation error within the ASTER DEMs analysis of these results and identification of surface elevation change values between 2000-2004 and 2004-2005, and comparison with previous studies cannot be completed. Consequently, this study has highlighted that the use of multi-temporal ASTER DEMs to measure elevation changes is clearly very sensitive to the actual quality of the DEM being used (Racoviteanu et al., 2007). Therefore, DEMs with both high spatial and vertical resolutions are a necessity, to identify differences and the spatial distribution of surface changes accurately (Etzelmuller, 2000).

7.5. Surface Velocity

Glacier surface velocity was estimated so that values could be compared to previous studies velocity calculations (Diolaiuti et al., 2005), to assess whether a reduction or increase in velocities has occurred.
Optical ASTER imagery was used along with an ASTER DEM (for orthophoto generation of the ASTER optical imagery) in a change detection approach. This approach is based on tracking the displacement of features in each satellite image over time (Kaab, 2005). Once surface velocities are calculated, the implication of this on the movement/re-distribution of the debris cover and the impact of this on other areas of this study (debris thickness estimation, surface elevation changes) can be investigated.

Figure 7.9: Surface velocities between 2004-2005. The Miage Glacier outline is shown to illustrate areas of zero velocity on the valley sides.

Surface velocity was estimated (Figure 7.9) using a feature tracking method on orthorectified ASTER bands 1-3 from 2004 and 2005 by Dr Adrian Luckman of Swansea University. Orthorectified images are used to minimise geometrical errors when satellite images used are from
different tracks. Therefore, the images are orthorectified to the same DEM to project them to the
same reference system and co-register the images and ensure geolocational accuracy (Leprince et
al., 2007; Scherler et al., 2008). This ensures that the outputs from the feature tracking processes
are as accurate as possible, as they are very sensitive to registration errors, because any changes in
the location of a feature will be identified as a movement (and related to velocity).

The impact of registration errors is clearly highlighted by figure 7.9 with a number of areas
experiencing either unexpected movement or movements much greater than predicted. The first
area of concern on the 2004-2005 velocity map (Figure 7.9) is an area located on the north east
flank of the glacier in the accumulation area, where an area of movement to the north east (from
the mid glacier region up the mountain sides) is identified, with velocities reaching around 40 m y
¹. This is unrealistic as the valley sides do not experience movement. Therefore, reasons for this
were investigated.

One explanation is the result of error in the registration between the two orthophoto causing
movement where it should not be found. After closer investigation of the location of this area it
was found to be the location of an area of very rapid back-wasting around a large moulin (which
grew rapidly from 2004-2008) which may also account for this anomalous movement. However,
fuller investigation is required to determine whether this is the cause. Errors are also present
where velocity values are identified on the mountain tops (where no movement occurred) (Figure
7.9). This occurs due to mis-registration of the two orthophotos, and due to errors within the
DEM used for the orthophoto generation of an image, which results in horizontal shifts in the
projection of the orthophoto (Kaab, 2005).
As reasons for these errors are known they could be excluded from analysis, however, as errors are present at these locations it makes analysis at other locations difficult as errors may also be present. In turn, this means comparisons to previous velocities cannot be completed as the reliability of the data is in question. Therefore, analysis of this data was not completed, however, as a reason for these errors is known (mis-registration errors due to DEM error) another velocity study could be completed using better quality DEMs to produce more reliable velocity estimates for analysis. This may also provide validation for these results or confirm that they are inaccurate.

7.6. Summary

This chapter focused upon two inter-related issues using ASTER DEM products. First, the estimation of surface elevation changes, and second, the estimation of surface velocities. Limitations in extracting elevation changes were encountered, relating to the quality of ASTER DEM products, due to the impact of snow and cloud upon their accuracy, which in turn limited the analysis of elevation changes and surface velocity estimations.

The measurement of elevation changes is a vital application as it provides information relating to the mass balance status of a glacier through positive or negative increases in elevation and, in turn, impacts of climate change upon this can be investigated. However, due to problems with the accuracy of ASTER DEMs a reliable estimate of elevation changes between 2000-2006 could not be completed. Therefore, the limitations of ASTER DEM products in similar study sites comprising of steep and varied terrain are highlighted. In turn, this identifies the requirement for DEMs that are not impacted by the terrain of a study site, because this and other previous studies suggest ASTER DEM is not fully suitable for glacier surface elevation change studies. Problems with the ASTER DEMs also affected the calculation of surface velocity measurements, with
movement identified on mountain tops due to orthophoto registration errors (a result of DEM errors). Consequently, analysis and comparison to previous field based studies of velocity, and identification of any changes in velocity rates over time, could not be completed.

Overall, this chapter has highlighted the limitations of ASTER DEMs for surface elevation change and velocity analysis. Therefore, this demonstrates the need for a reliable remotely sensed DEM, which can be obtained at a high temporal frequency for incorporation into methods which could be utilised by global monitoring programs such as GLIMS for analysing the impact of climatic variations on surface elevation and velocity changes on a greater number of debris-covered glaciers at a repeated timescale, greater than that which would be possible with fieldwork alone. In turn, this will enable increasingly detailed pictures of global changes in surface elevation and velocities on debris-covered glaciers to be identified. However, until this is available analysis will be restricted to those glaciers where more reliable airborne derived DEMs are available.
CHAPTER 8: MAPPING OF LITHOLOGY USING REMOTELY SENSED DATA; DEVELOPMENT AND TESTING USING ASTER IMAGERY

Chapter aim: To map the distribution of rock types in the supraglacial debris cover using spectral information derived from laboratory and satellite based remote sensing techniques.

8.1. Introduction

Detailed geological information is usually obtained from in-situ investigations (Jensen, 2000). However, as discussed previously, access to many high mountain regions is difficult and so lithological data in these regions is often hard to obtain. This is true of many debris-covered glaciers, and consequently limited data is available on the lithological composition of these debris layers. Knowledge of debris rock type is important as different rock types have different albedo, emissivity, and thermal conductivity values (along with other thermal properties) which can have a direct impact upon sub-debris ice melt rates.

Furthermore, the different emissivity values of different rock types may affect the retrieval of surface temperature from thermal satellite imagery, which will in turn have an impact upon debris thickness estimated using the energy balance model identified in chapter 5. Therefore, the implications of variable lithology of a debris layer (and possible variation in emissivity) on surface temperature estimates requires investigation, because currently the LP DAAC (who process the ASTER data) uses an average emissivity for each pixel to obtain surface temperature.

Remote sensing can potentially provide a solution to this problem, as it is possible to discriminate between different rock types in remotely sensed imagery due to their different spectral signatures,
resulting from differing chemical and mineral compositions (Jensen, 2000; Krezhova et al., 2000; Strasen et al., 2009; Amer et al., in press). These spectral differences allow different rock units to be mapped using satellite imagery. ASTER imagery has increasingly been used in lithology mapping studies due to its wide spectral coverage and high spatial resolution, which enhances its ability to discriminate different rock units compared to other systems such as Landsat TM (e.g. Rowan and Mars 2003; Ninomiya et al., 2005; Gad and Kusky, 2007; Strasen et al., 2009).

Consequently, knowing the rock emissivities present in a glaciers debris cover will enable the refinement of the energy balance model for estimating debris thickness, which will in turn, produce more reliable debris thickness estimates. Especially since fuzzy/soft classifications could identify the range of emissivity values present in each pixel (through the identification of different rock types), and resulting surface temperature calculations (and resulting debris thickness estimations) could take this into account.

Unlike previous studies, which have tended to use ratio images or a principle components analysis (PCA) to enhance the spectral contrasts between various rock types in a region (e.g. Rowan and Mars, 2003; Gad and Kusky, 2007; Strasen et al., 2009; Amer et al., in press), this study will combine methods that have previously been applied to hyperspectral imagery to undertake a supervised classification of ASTER data to identify the different rock types present across the entire debris-covered portion of the Miage Glacier. Along with analysis into emissivity variations of a debris cover due to the presence of different rock types and the implications on surface temperature retrieval from variable emissivity values. Therefore, being able to identify different rock types present within a debris layer unlocks the potential of ASTER data in determining
surface emissivity variations on debris-covered glaciers, meaning a more accurate (and spatially variable) surface temperature estimate can be produced.

8.2. Methodology

A number of different processes were completed during the generation of a rock lithology map, first rock samples collected in the field were analysed using a laboratory spectroradiometer. A cluster analysis was then performed on this data, which clusters the samples together based on their spectral signatures. Next these clusters were analysed to determine whether the clusters identified did represent different rock types, or whether due to spectral similarities some rock types could not be discriminated clearly. Secondly, using both the cluster analysis data and field data for training data, a supervised classification of the ASTER image was completed. Finally, once a supervised classification was undertaken, an unsupervised approach was also applied to assess whether a map lithology map could be produced without field data.

8.2.1. Rock unit spectral reflectance

Rock samples were collected in June 2007 from 17 locations across the Miage Glacier surface, with at least three different sample rocks taken from each location and 61 samples collected in total (locations of these samples can be found in section 4.4.7). Lithological identification of all of the samples was also completed, and details can be found in table 8.2. The spectral reflectance of each sample was measured in the laboratory with the spectroradiometer, with spectra collected from both the top and bottom of each rock. The spectroradiometer used was an ASD Fieldspec Pro Spectroradiometer, which has a spectral range of 350-2500 nm and spectral sampling of 1.4 nm at 350 - 1050 and 2 nm at 1000-2500 nm. The spectroradiometer foreoptics were fitted with a fibre optic contact probe that provided its own illumination, and each rock spectra collected by
holding the rock sample against the contact probe. If any of the rocks had distinctive differences in appearance, such as an area which was much darker in colour or composed of a different rock material, an additional measurement was also taken at this point to see the impact upon the spectral signature of the rest of the rock.

Figure 8.1: Rock sample spectral signatures a) little variation between the top and bottom (site 1C), b) and clear variation on each side of the rock (site 10E).
The spectral signatures from the top and bottom of the rock were then assessed to identify if any significant variation occurred. When little variation between the top and bottom sides occurred, (e.g. Figure 8.1 a) an average was generated from these two spectra, to generate an average spectra for the rock sample as a whole. If clear differences were apparent then an average was not generated and each spectra treated as a separate sample (e.g. Figure 8.1 b). Samples that showed such differences are representative of the variation apparent in each of the debris rock units. Therefore, it was important to retain these measurements to capture all of the spectral variation.

8.2.2. Cluster analysis

Cluster analysis was performed on the spectra in an approach similar to that of Krezhova et al., (2007). This was completed to determine whether the rock samples could be clustered based on their spectral similarities and differences. In turn, this aids subsequent supervised and unsupervised classification by determining the number of spectral clusters (and hence likely rock types) which could be discriminated spectrally.

Cluster analysis determines the number of groups/clusters in the data based on similarity in reflectance patterns. The distance (Euclidean distance) between these similarities is then used as the basis for their grouping. This process starts with the two most similar samples being paired together, and with other pairs added to this group if they are also similar. If pairs are not similar, a new group is created to which other paired samples can be compared if they are similar, or another new group created. This process continues until all of the samples have been grouped (Wheeler et al., 2006).
The cluster output identifies which samples can be grouped together based on their similarity (and the level of their similarity) and is shown diagrammatically through a dendrogram. The dendrogram shows the samples in their groups and at what level of similarity each sample was added. The greater the distance between each of the cluster centres, the lower the level of similarity between the samples in each class (Wheeler et al., 2006).

8.2.3. Supervised classification

Once the cluster analysis was performed and the spectral grouping of the different rock samples completed, a supervised classification was performed, with the cluster analysis providing a guide to how many different spectral clusters were present on the Miage Glacier (in conjunction with field identification of different rock groups). Supervised classification was performed on a multispectral ASTER image from 01/08/05 with all 14 wavebands and training sites identified during fieldwork in 2007. These sites identified the locations of the main rock types on the middle reaches of the glacier (Figure 4.11), although most areas are a mixture of different rock types, making identification and mapping difficult. As a result, a visual assessment in the field noted the dominant type of rock present in each location.

A maximum likelihood classification algorithm was selected which assigned pixels to classes based on their probability of belonging to that class. Six input classes (Figure 8.3) were selected based on the study site sketch map (Figure 4.11), and a rock lithology map from 2002 (Figure 8.2). However, initial classification outputs contained detail that included the valley sides, which were not of interest, and so an additional class was added with training sites generated from areas on the valley sides. The supervised classification was then repeated, and the classified image identified the slopes as one class, which were then removed from further analysis.
**Figure 8.2:** Map of geological units in surface debris on Miage Glacier (Deline, 2002), where L0 Rusted debris, L1 Mont Blanc granite (L1a Coarse grained, L1b medium grained, L1c granite with feldspars), L2 Micro granite, L3 Gneiss, L4 Schists (L4a Chloroschists, L4b Black crystal schists, L4c Ochreous schists, L4d White schists) L5 Ardoisier schist, L6 Quartzites, L7 Amphibolite, L8 tectonic breccias.
Figure 8.3: Training sites identified from the 2007 field map of the location of major rock types
(R:2, G:3, B:4).

Table 8.1: Number of training sites and total number of pixels within these training sites for each
rock unit, and slope class

<table>
<thead>
<tr>
<th>Rock type</th>
<th>Number of training sites</th>
<th>Total number of pixels</th>
</tr>
</thead>
<tbody>
<tr>
<td>Gniess</td>
<td>3</td>
<td>137</td>
</tr>
<tr>
<td>Granite</td>
<td>5</td>
<td>246</td>
</tr>
<tr>
<td>Rusted debris</td>
<td>11</td>
<td>743</td>
</tr>
<tr>
<td>Schist</td>
<td>6</td>
<td>306</td>
</tr>
<tr>
<td>Slate</td>
<td>9</td>
<td>432</td>
</tr>
<tr>
<td>Tectonic Breccia</td>
<td>3</td>
<td>141</td>
</tr>
<tr>
<td>Slopes</td>
<td>8</td>
<td>9214</td>
</tr>
</tbody>
</table>

8.2.4. Validation Data

Due to the limited amount of field data on rock types and their location on the Miage Glacier, two
different accuracy assessment methods were applied as a large independent validation data set was
unavailable, which restricted the ways in which accuracy could be assessed. First, using a
confusion matrix the accuracy of the training data sites after classification was investigated.
However, because the confusion matrix provides only the accuracy against the training sites it is
not an indication of the overall accuracy of the supervised classification on the entire image.
Therefore, another form of validation was required to assess the accuracy of the model when
applied to the whole image. This second approach utilised the limited amount of field data variable on rock type (17 sites in 2007, 9 sites in 2009), with the rock type in the classified image compared to the actual rock type recorded in the field. Photographs taken in the field at known locations in previous years (2 transects in 2006 and 2007) were also utilised, because the rock types observed in these photographs could be compared to the rock type identified by the supervised classification.

8.2.5. Unsupervised classification

To test whether this method of generating rock type distribution on a debris-covered glacier could be applied to other debris-covered glaciers with no field data, an unsupervised classification approach was applied to the Miage Glacier, so that results could be compared to those obtained using the supervised classification approach. Several different unsupervised options were tested with results analysed visually. The approach which performed best was deemed to be the ISODATA algorithm with 4 iterations, and a change threshold of 0.99%.

8.3. Results

8.3.1. Rock samples

As previously mentioned in section 8.2.1, 61 rock samples were collected in the field in 2007. Each of these samples were classified, with any distinctive features noted, and are shown in Table 8.2. These results show that, despite similar rocks being found clustered together at some sites, at others the range of rocks present are highly varied. Therefore, this highlights that although the debris layer comprises of distinctive areas comprising of a dominant rock type, which are highlighted by the field map in 2007 and rock lithology map from 2002, the variability at other locations (and within some of these distinctive areas) is high. This variability is due to the mixing
of input sources consisting of different rock types during its movement on the glaciers surface.

Therefore, variability will increase with distance down glacier due to both a greater number of
different debris input sources and amount of debris mixing with increasing distance down glacier.

**Table 8.2: Classified rock samples collected in 2007**

<table>
<thead>
<tr>
<th>Site</th>
<th>Rock type</th>
<th>Distinctive features</th>
<th>Site</th>
<th>Rock type</th>
<th>Distinctive features</th>
</tr>
</thead>
<tbody>
<tr>
<td>1A</td>
<td>Mica Schist</td>
<td>Muscovite rich</td>
<td>9B</td>
<td>Amphibole gneiss</td>
<td></td>
</tr>
<tr>
<td>1B</td>
<td>Quartzite</td>
<td>Granular</td>
<td>9C</td>
<td>Amphibole gneiss</td>
<td></td>
</tr>
<tr>
<td>1C</td>
<td>Quartzite</td>
<td></td>
<td>10A</td>
<td>Amphibole Gneiss</td>
<td></td>
</tr>
<tr>
<td>1D</td>
<td>Gneiss</td>
<td>Mica rich one side, Quartz Biotite rich other</td>
<td>10B</td>
<td>Schist Quartz Mica</td>
<td>Rusted/tarnished</td>
</tr>
<tr>
<td>2A</td>
<td>Quartz Biotite Gneiss</td>
<td>Similar to 1D</td>
<td>10C</td>
<td>Phyllite</td>
<td></td>
</tr>
<tr>
<td>2B</td>
<td>Amphibole Gneiss</td>
<td></td>
<td>10D</td>
<td>Biotite Mica Schist</td>
<td></td>
</tr>
<tr>
<td>2C</td>
<td>Muscovite Mica Schist</td>
<td>Similar to 1A</td>
<td>10E</td>
<td>Quartz Biotite Schist</td>
<td></td>
</tr>
<tr>
<td>2D</td>
<td>Rich fault rock</td>
<td>Quartz grains</td>
<td>10F</td>
<td>Quartz Mica Schist</td>
<td>Rusted</td>
</tr>
<tr>
<td>3A</td>
<td>Phyllite</td>
<td>Mica rich Biotite</td>
<td>11A</td>
<td>Slate</td>
<td></td>
</tr>
<tr>
<td>3B</td>
<td>Mica schist</td>
<td></td>
<td>11B</td>
<td>Slate</td>
<td></td>
</tr>
<tr>
<td>3C</td>
<td>Biotite Mica Schist</td>
<td>High Biotite</td>
<td>11C</td>
<td>Slate</td>
<td></td>
</tr>
<tr>
<td>4A</td>
<td>Quartz mica schist</td>
<td></td>
<td>12A</td>
<td>Quartzite</td>
<td></td>
</tr>
<tr>
<td>4B</td>
<td>Biotite schist</td>
<td></td>
<td>12B</td>
<td>Biotite schist</td>
<td></td>
</tr>
<tr>
<td>4C</td>
<td>Sheared Quartz</td>
<td>Veined</td>
<td>12C</td>
<td>Quartz mica schist</td>
<td></td>
</tr>
<tr>
<td>4D</td>
<td>Biotite Mica Schist</td>
<td></td>
<td>12D</td>
<td>Biotite schist</td>
<td></td>
</tr>
<tr>
<td>5A</td>
<td>Chlorite schist</td>
<td></td>
<td>12E</td>
<td>Quartz mica schist</td>
<td></td>
</tr>
<tr>
<td>5B</td>
<td>Chlorite schist</td>
<td></td>
<td>12F</td>
<td>Gneiss</td>
<td>Quartz and Feldspar</td>
</tr>
<tr>
<td>5C</td>
<td>Chlorite schist</td>
<td></td>
<td>13A</td>
<td>Quartzite</td>
<td></td>
</tr>
<tr>
<td>6A</td>
<td>Quartz Biotite Gneiss</td>
<td></td>
<td>13B</td>
<td>Biotite gneiss</td>
<td></td>
</tr>
<tr>
<td>6B</td>
<td>Quartz Biotite Gneiss</td>
<td></td>
<td>13C</td>
<td>Gneiss</td>
<td>Quartz rich</td>
</tr>
<tr>
<td>6C</td>
<td>Quartz Amphibolites</td>
<td></td>
<td>13D</td>
<td>Amphibole Gneiss</td>
<td></td>
</tr>
<tr>
<td>6D</td>
<td>Amphibole gneiss</td>
<td></td>
<td>13E</td>
<td>Quartz vein rock</td>
<td>Quartz crystals</td>
</tr>
<tr>
<td>7A</td>
<td>Granite</td>
<td>Rusted</td>
<td>14A</td>
<td>Amphibole Gneiss</td>
<td></td>
</tr>
<tr>
<td>7B</td>
<td>Amphibole</td>
<td></td>
<td>14B</td>
<td>Chlorite Schist</td>
<td>Green/powdery</td>
</tr>
<tr>
<td>7C</td>
<td>Quartz chlorite</td>
<td>Quartz vein rock</td>
<td>14C</td>
<td>Amphibole Gneiss</td>
<td>Amphiboles Quartz</td>
</tr>
<tr>
<td>8A</td>
<td>Quartz gneiss</td>
<td>Crystals in Amphibole</td>
<td>14D</td>
<td>Quartz Mica Schist</td>
<td></td>
</tr>
<tr>
<td>8B</td>
<td>Quartzite</td>
<td></td>
<td>14E</td>
<td>Amphibole Gneiss</td>
<td>Green crystals</td>
</tr>
<tr>
<td>8C</td>
<td>Quartz Biotite schist</td>
<td></td>
<td>15A</td>
<td>Chlorite Schist</td>
<td></td>
</tr>
<tr>
<td>8D</td>
<td>Amphibole</td>
<td></td>
<td>15B</td>
<td>Quartz rich Gneiss</td>
<td></td>
</tr>
<tr>
<td>9A</td>
<td>Amphibole gneiss</td>
<td>Band of quartz</td>
<td>15C</td>
<td>Amphibole Gneiss</td>
<td></td>
</tr>
<tr>
<td>9B</td>
<td>Amphibole gneiss</td>
<td></td>
<td>15D</td>
<td>Amphibole gneiss</td>
<td></td>
</tr>
</tbody>
</table>
Example spectral responses of rock samples representing each of the 6 major rock types present on the Miage Glacier (gneiss, granite, schist, tectonic breccia, slate, and rusted debris) are shown in Figures 8.4-8.9. These spectral response curves highlight the reflectance and absorption feature differences between each of the rock types, making the discrimination between them possible.

**Figure 8.4:** a) sample of schist collected at site 12D, b) spectral response of sample 12D.

**Figure 8.5:** a) sample of gneiss collected at site 14C, b) spectral response of sample 14C.
Figure 8.6:  
a) sample of rusted debris collected at site 10B, b) spectral response of sample 10B.

Figure 8.7:  
a) sample of slate collected at site 11A, b) spectral response of sample 11A.

Figure 8.8:  
a) sample of granite collected at site 7A, b) spectral response of sample 7A.
8.3.2. Cluster analysis results

Cluster analysis was completed using 2 different data sets:

1. All wavelengths collected by the spectroradiometer

This would provide the most spectral detail, increasing spectral separability (or similarity) of the rock samples due to the high spectral resolution of the spectroradiometer.

2. Spectroradiometer wavelengths corresponding to bands 1-9 of the ASTER sensor (spectroradiometer did not record data in the thermal infrared region).

This provides a more realistic assessment of the capability of the ASTER sensor to spectrally discriminate between the different rock types present on the Miage Glacier or whether a higher spectral resolution is required.

The midpoint of each ASTER waveband was identified and the corresponding spectroradiometer wavelength used to simulate the response from ASTER. When each of the dendrogram outputs were compared it was clear that the same number of clusters (5) was identified when both all
spectroradiometer bands (Figure 8.10a) and spectroradiometer bands corresponding to ASTER wavebands 1-9 were used (Figure 8.10b).

---

**Figure 8.10:** ‘Cut’ dendrogram output for all collected rock samples, where each colour represents a new class, a) using their spectral response at all wavelengths of the spectroradiometer, b) ASTER bands 1-9.

The clusters were colour coded during the ‘cut’ process to aid cluster identification. The dendrogram was cut in 5 places, with the smallest distance possible between each of the cluster centres, with the distance identified by the horizontal lines on the dendrogram. If the dendrogram had been cut at fewer locations, the distance between the cluster centres would have been much greater, and therefore, the lower level of similarity between the samples in each class.
Consequently, the cluster process is not based on how well rock types fit into the clusters, instead the clusters identify the level of similarity (or difference) between the different rock samples.

However, the percentage of sites which appear in the same group using both the spectroradiometer and ASTER (Table 8.3) are different, with only the orange class having greater than 50% (63%) of samples grouped within the same class independent of the sensor used. The red class had no sites which appeared within the same clusters, and in the blue and green classes only a small percentage of sites appear in the same class (17% and 25% respectively).

<table>
<thead>
<tr>
<th>Class</th>
<th>% sites which appear in the same class</th>
</tr>
</thead>
<tbody>
<tr>
<td>Orange</td>
<td>63 (25/40)</td>
</tr>
<tr>
<td>Purple</td>
<td>40 (2/5)</td>
</tr>
<tr>
<td>Green</td>
<td>25 (2/8)</td>
</tr>
<tr>
<td>Red</td>
<td>0 (0/3)</td>
</tr>
<tr>
<td>Blue</td>
<td>17 (1/6)</td>
</tr>
</tbody>
</table>

Variations in the groupings occur due to the different range of wavelengths used, meaning that some samples which were easily discriminated when all bands of the spectroradiometer were used, may be more difficult to discriminate with a reduced spectral resolution when the ASTER satellite is used. These difficulties arise due to the mixed mineralogy of the rocks at the Miage Glacier, which makes them difficult to separate (or cluster) spectrally. Therefore, the clusters identified by the spectroradiometer with all wavebands provide the most accurate grouping of the field samples due to the spectroradiometers high spectral resolution.
The clusters identified using the spectroradiometer wavelengths corresponding to bands 1-9 of the ASTER sensor are not able to cluster the rock samples as well due to the reduction in spectral resolution compared to the spectroradiometer. This makes the discrimination of rock samples more difficult as some exhibit key features in areas of the spectrum not covered by the ASTER bands. If the spectroradiometer had recorded information in the thermal region a more reliable cluster analysis would be produced using the bands corresponding to the ASTER satellite, as the inclusion of the spectroradiometer data at bands corresponding to the ASTER thermal bands would increase the spectral range from 9 bands to 14. This increase would enable greater spectral separation or grouping of the samples, because many display key spectral features in the thermal region.

The impact of this differential grouping will be apparent after the rock classification using ASTER data (which also utilises the thermal bands), by identifying whether the different rock types present on the Miage Glacier (and their distribution patterns) can be correctly discriminated.

**8.3.3. Supervised classification results**

The results of the supervised classification (Figure 8.11) showed a reasonable accuracy (81%) between the training sites and the classified data (Table 8.4a) (kappa coefficient of 0.76). Referring to the producer accuracies (Table 8.4 b), the greatest producers accuracies are obtained for gneiss (96%), granite (92%), and tectonic breccia (91%), highlighting the increased separability of these classes from others in the image.
Figure 8.11: Supervised classification outputs with the slope class removed.

Table 8.4: a) confusion matrix and accuracy of supervised classification, b) producer’s and user’s accuracies

a)

<table>
<thead>
<tr>
<th>Class</th>
<th>Gneiss</th>
<th>Granite</th>
<th>Rusted debris</th>
<th>Schist</th>
<th>Slate</th>
<th>Tectonic breccia</th>
<th>Total</th>
</tr>
</thead>
<tbody>
<tr>
<td>Unclassified</td>
<td>0</td>
<td>2</td>
<td>28</td>
<td>11</td>
<td>11</td>
<td>0</td>
<td>52</td>
</tr>
<tr>
<td>Gneiss</td>
<td>132</td>
<td>0</td>
<td>0</td>
<td>1</td>
<td>1</td>
<td>0</td>
<td>134</td>
</tr>
<tr>
<td>Granite</td>
<td>0</td>
<td>227</td>
<td>70</td>
<td>14</td>
<td>34</td>
<td>0</td>
<td>345</td>
</tr>
<tr>
<td>Rusted debris</td>
<td>1</td>
<td>3</td>
<td>565</td>
<td>4</td>
<td>18</td>
<td>3</td>
<td>594</td>
</tr>
<tr>
<td>Schist</td>
<td>0</td>
<td>9</td>
<td>25</td>
<td>247</td>
<td>32</td>
<td>3</td>
<td>316</td>
</tr>
<tr>
<td>Slate</td>
<td>4</td>
<td>5</td>
<td>41</td>
<td>27</td>
<td>326</td>
<td>7</td>
<td>410</td>
</tr>
<tr>
<td>Tectonic breccia</td>
<td>0</td>
<td>0</td>
<td>14</td>
<td>2</td>
<td>10</td>
<td>128</td>
<td>154</td>
</tr>
<tr>
<td><strong>Total</strong></td>
<td><strong>137</strong></td>
<td><strong>246</strong></td>
<td><strong>743</strong></td>
<td><strong>306</strong></td>
<td><strong>432</strong></td>
<td><strong>141</strong></td>
<td><strong>2005</strong></td>
</tr>
</tbody>
</table>

Accuracy = (1625/2005) 81%, Kappa Coefficient = 0.76

b)

<table>
<thead>
<tr>
<th>Class</th>
<th>Producer’s Accuracy % (omission errors)</th>
<th>User’s Accuracy % (commission errors)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Gneiss</td>
<td>96 (132/137)</td>
<td>99 (132/134)</td>
</tr>
<tr>
<td>Granite</td>
<td>92 (227/246)</td>
<td>66 (227/345)</td>
</tr>
<tr>
<td>Rusted debris</td>
<td>76 (565/743)</td>
<td>95 (565/594)</td>
</tr>
<tr>
<td>Schist</td>
<td>81 (247/306)</td>
<td>78 (247/316)</td>
</tr>
<tr>
<td>Slate</td>
<td>75 (326/432)</td>
<td>80 (326/410)</td>
</tr>
<tr>
<td>Tectonic Breccia</td>
<td>91 (128/141)</td>
<td>83 (128/154)</td>
</tr>
</tbody>
</table>
Lower producer accuracies (Table 8.4b) are obtained for the rusted debris (76%) and slate (75%) classes, indicating a lower spectral separability. In turn, this resulted in a greater number of pixels being incorrectly classified as different rock types, such as the rusted debris class, which included a number of granite (70), slate (41) and tectonic breccia (14) pixels. The slate class was the only other class to have a large number of pixels incorrectly identified as another rock type, with most pixels being incorrectly classified as granite (34), followed by schist (32), and rusted debris (18).

This confirms observations that these classes (especially slate and granite) are spectrally similar (e.g. Shukla et al., 2009). The user’s accuracies (Table 8.4b) have a greater spread of values than the producer’s accuracies, with the lowest value obtained for granite (66%), where 70 pixels are incorrectly classified as rusted debris and 34 as slate. The highest accuracies were obtained for gneiss (99%) and rusted debris (95%), highlighting a high spectral separability of these from the other rock types. As all of the producer and user accuracies were greater than 60% and a high overall accuracy value for the confusion matrix and kappa coefficient was obtained, it indicates that the supervised classification of rock type has been applied successfully to the training areas.

The greatest source of error occurred due to spectral similarity between the classes, and may be a result of the training site areas not being completely homogenous, due in part to the variability of rock types over a small spatial area in the Miage Glacier. Compounding this is the spatial resolution of the ASTER image which will contain more than one rock type within each 90 m x 90 m pixel. Therefore, in some cases the spectral signature of the training sites is a mixture of different rock types, which along with the spectral similarity of some rock types, explains the presence of errors of omission and commission. Overall, therefore, table 8.4 suggests that coarse
grained crystalline lithotypes are most easily classified/identified on this basis, compared to finer
grained sizes, which are not as spectrally separable.

8.3.4. Comparison to field data

8.3.4.1. 2007 and 2009 field data

The observations of rock type made in the field were compared to those obtained with the
supervised classification. Some differences are apparent, where the rock classification has not
correctly identified the rock type(s) present in the field (Table 8.5). Despite these differences at
some sites, at least one of these rock types is successfully identified by the classification outputs.
At site 1 (2009), for example, where both rusted debris and granite is identified, or sites 18-19
(2007) where gneiss is identified along with schist (Table 8.5). Some of the validation sites were
located on the boundary between two different rock types and, in this case the rock type identified
in the field did correspond to one of the two rock types found at the validation point location (e.g.
sites 3 and 10). Also, at site 8 (2009), both of the two rock types identified in the field were
identified by the model, as the validation point was on the boundary of those two rock types.

This highlights that at some locations there is a good accuracy between the classified image and
the field data. At some locations, however, different rock types were identified by the
classification compared to those in the field, indicating that the comparison to the field data from
2007 and 2009 does not fully support the successful validation of the classified rock map.
However, despite these differences, it does not necessarily mean the classified map is invalid, as
reasons for the difference between the values identified in the field to those in the map can be
identified. The main reason being the limitations of using point sampling of rock types for
validating 90 x 90 m pixel areas. Also the lithological units will have been transported by ice
flow between 2005 and 2007/2009, which may account for some of the differences between the 2005 map and rock types identified in the field. Supervised classification of ASTER data would, therefore, seem to be a promising tool for mapping the general lithology patterns on debris-covered glaciers. However, with a 90 x 90 m spatial resolution, the application of soft or fuzzy classification approaches may be beneficial and should be further investigated.

Table 8.5: 2007 and 2009 field samples of rock type against classification rock type for the corresponding pixel, if a sample point was located on the boundary of two rock types both rock types are identified

<table>
<thead>
<tr>
<th>Site</th>
<th>Rock types present</th>
<th>Classification rock type</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Schist Rusted</td>
<td>Gneiss</td>
</tr>
<tr>
<td>2</td>
<td>Gneiss/schist rusted</td>
<td>Gneiss</td>
</tr>
<tr>
<td>3</td>
<td>Gneiss/rusted</td>
<td>Gneiss/rusted</td>
</tr>
<tr>
<td>4</td>
<td>Gneiss/schist</td>
<td>Gneiss</td>
</tr>
<tr>
<td>5</td>
<td>Slate</td>
<td>Gneiss/rusted</td>
</tr>
<tr>
<td>6</td>
<td>Schist/gneiss rusted</td>
<td>Gneiss</td>
</tr>
<tr>
<td>7</td>
<td>Schist/rusted</td>
<td>Gneiss/rusted</td>
</tr>
<tr>
<td>8</td>
<td>Schist</td>
<td>Rusted</td>
</tr>
<tr>
<td>9</td>
<td>Schist</td>
<td>Rusted</td>
</tr>
<tr>
<td>10</td>
<td>Schist/rusted</td>
<td>Rusted/tectonic breccia</td>
</tr>
<tr>
<td>11</td>
<td>Schist</td>
<td>Rusted</td>
</tr>
<tr>
<td>12</td>
<td>Gneiss, schist</td>
<td>Slate</td>
</tr>
<tr>
<td>13</td>
<td>Gneiss, schist</td>
<td>Granite/rusted</td>
</tr>
<tr>
<td>14</td>
<td>Gneiss</td>
<td>Rusted/schist</td>
</tr>
<tr>
<td>15</td>
<td>Gneiss/schist</td>
<td>Schist/rusted</td>
</tr>
<tr>
<td>16</td>
<td>Schist/gneiss</td>
<td>Schist/gneiss</td>
</tr>
<tr>
<td>17</td>
<td>Schist/gneiss</td>
<td>Gneiss</td>
</tr>
<tr>
<td>18</td>
<td>Mixture, including rusted</td>
<td>Rusted debris</td>
</tr>
<tr>
<td>19</td>
<td>Mixture, but no rusted/granite</td>
<td>Rusted debris</td>
</tr>
<tr>
<td>20</td>
<td>Schist</td>
<td>Slate/rusted debris</td>
</tr>
<tr>
<td>21</td>
<td>Schist</td>
<td>Unclassified</td>
</tr>
<tr>
<td>22</td>
<td>Schist, some rusted</td>
<td>Slate</td>
</tr>
<tr>
<td>23</td>
<td>60% slate, 40% rusted</td>
<td>Rusted debris/slate</td>
</tr>
<tr>
<td>24</td>
<td>Schist, some rusted</td>
<td>Slate/schist</td>
</tr>
<tr>
<td>25</td>
<td>70% granite, 30% rusted</td>
<td>Rusted debris</td>
</tr>
<tr>
<td>26</td>
<td>Mixture</td>
<td>Schist/granite</td>
</tr>
<tr>
<td></td>
<td><strong>Total identified correctly</strong></td>
<td><strong>12/26</strong></td>
</tr>
</tbody>
</table>
8.3.4.2. Field photographs

Table 8.6: Rock types present at each field photograph site compared to rock type identified by the supervised classification

<table>
<thead>
<tr>
<th>Site</th>
<th>Rock type(s) present</th>
<th>Classification rock type</th>
</tr>
</thead>
<tbody>
<tr>
<td>1 - Lower transect 2006</td>
<td>Schist (mainly), Granite, Rusted debris, Slate, Gneiss</td>
<td>Schist</td>
</tr>
<tr>
<td>2 - Middle transect 2006</td>
<td>Granite/schist (mainly), Rusted debris, Slate, Gneiss</td>
<td>Granite</td>
</tr>
<tr>
<td>3 - Upper transect 2006</td>
<td>Tectonic breccia, Granite</td>
<td>Tectonic breccia</td>
</tr>
<tr>
<td>4 - Lower transect 2007</td>
<td>Granite/schist (mainly), Rusted debris, Slate, Gneiss</td>
<td>Schist</td>
</tr>
<tr>
<td>5 - Upper transect 2007</td>
<td>Granite (mainly), Tectonic breccia</td>
<td>Rusted debris (boundary with tectonic breccias)</td>
</tr>
</tbody>
</table>

Figure 8.12: Photograph of lower transect area in 2006 showing presence of schist and gneiss (along with other rock types), with corresponding extract from classified image suggesting schist dominates the same region.

Results from the comparison with photographs show good agreement between the rock types identified in the field and those identified by the image classification (Table 8.6), with the most dominant rock types identified at sites 1 – 4, and one of the less dominant rock types correctly identified at site 5. At site 1, for example, the classified image correctly identifies schist as the...
dominant rock type (Figure 8.12). Increased accuracy compared to the previous results occurs due to the use of a transect area rather than point location, and also the photographs were taken only 1 or 2 years after the classified image. This analysis suggests that due to the mixture of rock types present in most areas of the glacier, point sampling in the field is not a good method of identifying the dominant rock types within an area, particularly if used for image validation. An alternative method, such as the use of transects, where the percentage of each rock type present at regular intervals over a transect can be recorded may provide better comparison to the 90 x 90 m classified ASTER image.

8.3.5. Unsupervised classification results

Figure 8.13: Unsupervised classification of rock type distribution.
Different clusters on the unsupervised image (Figure 8.13) were identified as each of the six major rock types using the 2007 field map of the distribution of major rock types, and allocated a specific colour according to the previous scheme. When compared to rock types recorded in the field in 2007 and 2009 (Table 8.7), results suggest that the unsupervised map correctly identified at least one of the rock types at slightly more of the 2007 and 2009 field sample sites (16), compared to the supervised classification (Table 8.6; 13).

Table 8.7: 2007 Field samples of rock type against unsupervised classification rock type for the corresponding pixel, if a sample point was located on the boundary of two/three rock types all rock types are identified

<table>
<thead>
<tr>
<th>Site</th>
<th>Rock types present</th>
<th>Unsupervised classification rock type</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Schist/Rusted</td>
<td>Schist/Rusted</td>
</tr>
<tr>
<td>2</td>
<td>Gneiss/schist rusted</td>
<td>Gneiss/rusted/schist</td>
</tr>
<tr>
<td>3</td>
<td>Gneiss/rusted</td>
<td>Tectonic breccia/gneiss</td>
</tr>
<tr>
<td>4</td>
<td>Gneiss/schist</td>
<td>Gneiss</td>
</tr>
<tr>
<td>5</td>
<td>Slate</td>
<td>Tectonic breccia/granite</td>
</tr>
<tr>
<td>6</td>
<td>Schist/gneiss rusted</td>
<td>Gneiss/Rusted</td>
</tr>
<tr>
<td>7</td>
<td>Schist/rusted</td>
<td>Rusted/Gneiss</td>
</tr>
<tr>
<td>8</td>
<td>Schist</td>
<td>Rusted</td>
</tr>
<tr>
<td>9</td>
<td>Schist</td>
<td>Rusted/gneiss</td>
</tr>
<tr>
<td>10</td>
<td>Schist/rusted</td>
<td>Rusted/gneiss</td>
</tr>
<tr>
<td>11</td>
<td>Schist</td>
<td>Rusted/gneiss</td>
</tr>
<tr>
<td>12</td>
<td>Gneiss, schist</td>
<td>Rusted/Gneiss</td>
</tr>
<tr>
<td>13</td>
<td>Gneiss, schist</td>
<td>Rusted</td>
</tr>
<tr>
<td>14</td>
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<td>Rusted</td>
</tr>
<tr>
<td>15</td>
<td>Gneiss/schist</td>
<td>Rusted/gneiss/schist</td>
</tr>
<tr>
<td>16</td>
<td>Schist/gneiss</td>
<td>Gneiss</td>
</tr>
<tr>
<td>17</td>
<td>Schist/gneiss</td>
<td>Gneiss</td>
</tr>
<tr>
<td>18</td>
<td>Mixture, including rusted</td>
<td>Rusted/ schist/gneiss</td>
</tr>
<tr>
<td>19</td>
<td>Mixture, but no rusted/granite</td>
<td>Gneiss</td>
</tr>
<tr>
<td>20</td>
<td>Schist</td>
<td>Gneiss/tectonic breccia/rusted</td>
</tr>
<tr>
<td>21</td>
<td>Schist</td>
<td>Gneiss/rusted</td>
</tr>
<tr>
<td>22</td>
<td>Schist, some rusted</td>
<td>Gneiss/rusted</td>
</tr>
<tr>
<td>23</td>
<td>60% slate, 40% rusted</td>
<td>Rusted/schist</td>
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<tr>
<td>24</td>
<td>Schist, some rusted</td>
<td>Rusted/gneiss</td>
</tr>
<tr>
<td>25</td>
<td>70% granite, 30% rusted</td>
<td>Rusted</td>
</tr>
<tr>
<td>26</td>
<td>Mixture</td>
<td>Granite</td>
</tr>
</tbody>
</table>

Total identified correctly: 16/26
However, despite this slight increase in accuracy at some field sites, the classified map using the supervised approach does not identify as much detail of both the spatial distribution and variability of rock types present throughout the glacier tongue (especially of slate and gneiss). This reduction in detail is demonstrated by the classified map showing a dominance of rusted debris throughout most of the glacier tongue, along with small patches of granite and gneiss. However, it does not identify any schist and/or slate on the glacier tongue, which is known to be present through fieldwork. Therefore, although a classified image showing the general distribution of rock types can be produced without field data through an unsupervised classification. The inclusion of field data through a supervised approach improves the retrieval of the overall spatial pattern and variability of debris types present, and is thus a more reliable method for estimating rock type distribution in glacier debris cover.

8.4. Implications

Being able to identify different rock types present within a debris layer unlocks the potential of ASTER data in determining surface emissivity variations on debris-covered glaciers, meaning a more accurate (and spatially variable) surface temperature estimate can be produced. Table 8.8 shows the variability in emissivity of different materials, and the impact of emissivity on estimated surface temperature, which clearly is considerable.

<table>
<thead>
<tr>
<th>Material</th>
<th>Emissivity $\varepsilon$</th>
<th>True kinetic temperature $K$</th>
<th>Radiant temperature $K$</th>
<th>Radiant temperature $^\circ C$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Blackbody</td>
<td>1.00</td>
<td>300</td>
<td>300</td>
<td>27.0</td>
</tr>
<tr>
<td>Distilled water</td>
<td>0.99</td>
<td>300</td>
<td>299.2</td>
<td>26.2</td>
</tr>
<tr>
<td>Rough basalt</td>
<td>0.95</td>
<td>300</td>
<td>296.2</td>
<td>23.2</td>
</tr>
<tr>
<td>Vegetation</td>
<td>0.98</td>
<td>300</td>
<td>298.5</td>
<td>25.5</td>
</tr>
<tr>
<td>Dry loam soil</td>
<td>0.92</td>
<td>300</td>
<td>293.8</td>
<td>20.8</td>
</tr>
</tbody>
</table>
Knowing the emissivity means that the surface temperature identified by the ASTER image (generated using an average emissivity value for each pixel) could be verified, and any variations analysed. Therefore, different emissivity values could be applied if the value used to originally estimate surface temperature is identified as being significantly different to the emissivity of rock type whose presence is indicated by the classification. It also provides a means of identifying emissivity and its variability over small areas and that this should be considered when estimating surface temperatures derived from satellite images using emissivity values.

Another key point is that two rocks of the same type next to one another, which should have the same (or very similar) surface temperatures, may have different temperatures when sensed by a remote sensing system due to differences in emissivity (Jensen, 2000). Consequently, although their ‘true kinetic temperatures’ are the same, the temperature that is recorded is different through the influence of variable emissivity. This also applies to two different surfaces which may be recorded as having the same radiant temperatures when in fact they are different (Table 8.8). Therefore, if variation in emissivity is not accounted for correctly, significant temperature under- or over-estimations in thermal imagery can occur (Jensen, 2000; Campbell, 2008).

Table 8.9 shows the impact of variability in emissivity on surface temperature calculations and resulting debris thickness estimations, with the surface temperature value calculated from the CRN1 sensor (which requires the user to input an emissivity value to calculate surface temperature - similarly to the ASTER AST08 image) (column 3). After surface temperature was calculated, debris thickness was estimated at the LWS site using the energy balance model and recorded meteorological details at the LWS on 01/08/05 at 10:40, where the measured debris thickness was 0.16 m.
Table 8.9: Example emissivity values for rock types found on the Miage Glacier (columns 1 and 2), and effect on surface temperature retrieval and calculated debris thickness values using the energy balance model, \(^1\)Jensen (2000), \(^2\)Oke (1987), \(^3\) Hunt (1989). Surface temperature values obtained by inputting the selected emissivity value into the CRN1 temperature calculation.

<table>
<thead>
<tr>
<th>Rock type</th>
<th>Emissivity</th>
<th>Surface temperature</th>
<th>Debris thickness (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Biotite Granite</td>
<td>0.80</td>
<td>37.01</td>
<td>0.35</td>
</tr>
<tr>
<td>Andesite</td>
<td>0.82</td>
<td>35.10</td>
<td>0.25</td>
</tr>
<tr>
<td>Quartz Monzonite</td>
<td>0.85</td>
<td>32.35</td>
<td>0.17</td>
</tr>
<tr>
<td>Granite</td>
<td>0.86</td>
<td>31.45</td>
<td>0.16</td>
</tr>
<tr>
<td>Andesite Breccia</td>
<td>0.88</td>
<td>29.71</td>
<td>0.14</td>
</tr>
<tr>
<td>Slate</td>
<td>0.90</td>
<td>28.02</td>
<td>0.12</td>
</tr>
<tr>
<td>Basalt</td>
<td>0.95</td>
<td>23.97</td>
<td>0.10</td>
</tr>
<tr>
<td><strong>Average</strong></td>
<td><strong>0.87</strong></td>
<td><strong>30.58</strong></td>
<td><strong>0.21</strong></td>
</tr>
</tbody>
</table>

These debris thickness estimates at the LWS using individual emissivity values were then compared to the debris thickness estimate (with the energy balance debris thickness model) at the LWS using an average emissivity value of all the rock types. Results highlight the impact of emissivity upon debris thickness estimations due to its impact upon calculated surface temperatures. For example, a difference in emissivity of only 0.02 (0.80 and 0.82) resulting in a temperature difference of \(-2 \degree{C}\), which in turn, has a significant impact on debris thickness estimations, resulting in a difference of 0.10 m. Also the use of an average emissivity value for calculating surface temperature produced an incorrect debris thickness estimate at the LWS. This illustrates the significant impact that the use of an incorrect emissivity value has in the retrieval of surface temperature, which in turn has implications upon the estimation of debris thickness from surface temperature images.

Different rock types also have different effective thermal conductivities. Therefore, the mapping of the spatial pattern of rock types could also potentially improve distributed modelling of sub-debris melt through variable thermal conductivity. Results from the supervised classification highlight the varied lithologies present on the Miage Glacier, and the number of different sources
of origin. Therefore, remote sensing of rock lithologies could be used to track material on debris covered glaciers and enable the identification of the source of debris. This means that if one type of debris is dominant, its source may provide an explanation for its dominance. If the debris source is known, the movement and redistribution of the debris cover could be identified, providing information on the reworking of the debris cover during transport from its source.

8.5. Summary

This chapter focussed upon the potential applicability of ASTER data to map the rock types present in the debris cover of the Miage Glacier. To achieve this, cluster analysis and a supervised classification were utilised, and results of rock type distribution were generally favourable when compared to field data. This provides support for the application of this method to other debris-covered glaciers, meaning the variation in emissivity at different debris-covered glaciers can be identified, and the impact upon surface temperatures (and calculated melt rates) determined. However, future work focussing on the comparison of surface temperatures generated using both individual emissivity values from rocks found within a 90 x 90 m area of an ASTER pixel, and the average emissivity value used by the ASTER team is required.

An unsupervised approach was also applied successfully to the ASTER image, although the spatial variability in rock types was not as clearly identified as the supervised classification. However, an unsupervised approach has potential applicability in regions of limited field data, to provide a map showing the general patterns of rock type of a debris layer, from which areas of interest could be identified for further investigation.
Limitations were identified through the application of point data for validating the classified rock map, highlighting the requirement for transect data measurements which are representative of an ASTER pixel both in location in the field (to ensure transects do not span two or more pixels) and spatial size. High levels of variability in rock type within an area were also identified from the field data, which generates difficulties in visually identifying the most dominant rock type over any 90 m x 90 m area. In turn, this has an impact upon the validation of the classification as the classified rock map may identify one of the rock types present but cannot account for the variability that occurs due to it being a hard rather than soft classification. Therefore, the potential for the application of a fuzzy or soft classification approach for mapping rock lithology is highlighted (e.g. Paasche and Eberle, 2009), through its potential ability to identify the different rock types present within a pixel. However, this requires further investigation to address whether it could be successfully applied.

The application of rock type mapping provides more sensitive and spatially representative data for the estimation of debris thickness when using the energy balance model, as different rock types can be considered in its application. This is an important issue to consider, because different rock types will warm at different rates, which in turn causes variations in $K$ and will, therefore, affect estimated debris thicknesses using the energy balance model. This analysis has also illustrated the utility of rock type mapping using a supervised classification approach to aid in the identification of the source of debris material present on a debris-covered glacier (if valley slopes are included in the analysis). Such information is useful in determining the amount of debris material input from different sources and how the debris cover is redistributed with movement down glacier.
9.1. Introduction

One key aim was identified at the start of this study, and focused upon testing the utility of visible/infrared satellite sensors with frequent repeat global coverage at a high temporal and spatial resolution, such as TERRA ASTER and Landsat (MSS, TM, and ETM+) for studying debris-covered glaciers. The extent to which this aim has been addressed is discussed below in 3 sections which investigate what was known before this study was completed, what is known now the study has been completed (how this study fills the gaps present in current research), and what future work needs to address. An assessment to the contribution of the findings of this study from a remote sensing and glaciological perspective is also included.

9.2. What was known prior to this study?

Investigation into previous work at the start of this project highlighted that compared to ‘clean’ glaciers significantly fewer studies had been completed on debris-covered glaciers. This was of key concern because the number of debris-covered glaciers is expected to increase under predicted global warming impacts. Therefore, this highlighted that increased focus on debris-covered glaciers was required, especially since the presence of a debris layer on a glaciers surface modifies the thermal regime of the surface, and in turn influences surface ablation rates. A key factor which affects the number of studies on both ‘clean’ and debris-covered glaciers is their accessibility, which restricts both the number of studies and the repeat frequency of these studies. Remote sensing was highlighted as a solution to this problem, through the provision of satellite data at any location with a high temporal resolution. A number of previous studies on debris-covered glaciers have taken advantage of remote sensing, and have looked at surface elevation
changes using DEMs, boundary mapping, velocity measurements, and debris cover extent variations using a manual and semi-automatic approach.

However, these studies were restricted in number, and few looked into the potential of remote sensing for estimating debris thickness. Those that have were limited to empirical methods which require large amounts of field data and are site specific. Studies on surface elevation changes and velocity measurements on debris covered glaciers had been previously applied. However, DEM errors had been encountered in many locations limiting their full application. Another area which has not received any previous focus is the mapping of rock types present within a supraglacial debris layer. This will have a key impact upon the spatial distribution of surface ablation, through the impact of different rock type emissivity and albedo values. These gaps in previous work, therefore, provided the focus for this study.

9.3. What is known now the study has been completed?

The focus of this study was to fill in the gaps present in the current research undertaken on debris-covered glaciers. These gaps were highlighted by the objectives and key research questions stated at the start of this project. Therefore, research undertaken in each of the key research chapters was specifically targeted to address these issues. The key contributions of each of these chapters to the existing research on debris-covered glacier monitoring are highlighted below.

**Objective 1:** To develop a method to estimate debris thickness from ASTER thermal-band imagery using a physically-based energy balance model.

Chapter 5 demonstrated that debris thickness can be estimated for a supraglacial debris cover using a physically-based energy balance approach, which utilised meteorological variables
measured at the site, and surface temperature from an ASTER image. Also, compared to empirical approaches which are dependant on extensive field measurements of debris thickness and surface temperature that are site specific, this method is potentially transferable as it does not require these extensive measurements. Modelled or remote meteorological data from can be used from years when in situ AWS data are not available, and NCEP/NCAR (National Centres for Environmental Prediction-National Centre for Atmospheric research) reanalysis data could be used at remote sites, as it can be obtained globally at different elevations from NOAA (National Oceanic and Atmospheric Administration).

To test the models transferability, application to another glacier needs to be completed, including application to locations with different topography and climatic zones. Otherwise transferability would only be supported for similar glaciers and, therefore, would not assess the full potential of the energy balance debris thickness model on a global scale. If application to another glacier is successful, it will demonstrate the physical model’s potential towards the global monitoring of debris-cover. However, a general issue identified was that the more physically-based a model becomes the more the model is affected by quality of input data used, such as a DEM in this case. Therefore, a more physically-based model which considers slope and aspect would include greater detail with regard to the input parameters used and, therefore, produce more reliable debris thickness estimates. To ensure a greater level of reliability, the model requires input data of significantly better quality and fine spatial resolution. Such data are unlikely to be available for most glaciers and hence a partly physical model is likely to be the best solution until high quality data are routinely available.

**Objective 2:** To use TERRA ASTER and Landsat data (MSS, TM, ETM+) to map changes in debris cover extent over time, using both manual and semi-automatic approaches.
Chapter 6 highlighted that the semi-automatic method of Paul et al., (2004) can be successfully applied to ASTER imagery, following its initial development utilising Landsat TM data. However, a problem with the slope threshold utilised in the semi-automatic method of Paul et al., (2004) is clearly highlighted. A value of 24° is not steep enough to include the steep medial moraines of the Miage Glacier, leading to misclassification of debris-covered areas (Foster, 2009). A possible refinement is to apply manual correction to these areas and although this would increase processing time, it would produce more reliable debris extent results. Despite an increased processing time it is still quicker than field observations or manual digitising and with a much higher temporal frequency. This misclassification of areas of debris-covered ice on the medial moraine (due to the critical slope angle used) also raises issues with the quality of the DEM applied in the semi-automatic method. To ensure correct debris-extent estimates, a good quality DEM from a year corresponding to the multispectral imagery is required. Otherwise a poor quality DEM will generate incorrect slope estimates for application into the model. These errors will result in a debris extent map which incorrectly classifies debris-covered or debris free ice.

It became clear following the completion of debris change analysis that debris-cover on the Miage Glacier has not significantly increased between 1990-2004, and therefore, the Glacier became extensively debris-covered prior to this date. Also, the presence and location of sporadic or patchy debris in the upper accumulation zone could not be identified using either method on ASTER and Landsat TM data. The presence and ability to distinguish this patchy or sporadic debris is of key importance in mass balance studies as ablation will be maximised at these locations. To address this problem, finer spatial resolution imagery could be applied at certain locations (areas of interest, such as the debris-ice margins) to improve this accuracy. Examples include Quickbird
imagery which has a spatial resolution of 0.61 m (Panchromatic) and 2.4 m (multispectral), and IKONOS data at 0.84 m (Panchromatic) and 3.2 m (multispectral).

**Objective 3:** To use ASTER DEMs and visible imagery to monitor both changes in surface elevation over time, and estimate surface velocities using feature tracking methods.

Issues of ASTER DEM accuracy in areas of steep and varied terrain were highlighted during the investigation of surface elevation changes and surface velocity variations over time in chapter 7. These errors made it impossible to accurately determine elevation or velocity changes and, therefore, an analysis of the results could not be performed. In turn, this highlights the need for an improvement of ASTER DEM accuracy in these regions, and requirement of a DEM with a temporal resolution similar to ASTER but with significantly better accuracies. Therefore, application of ASTER DEM products for surface elevation and velocity change identification in this project has been severely limited.

**Objective 4:** To map the distribution of rock types in the supraglacial debris cover using spectral information derived from laboratory and satellite based remote sensing techniques.

Chapter 8 described the generation of a map of rock types in the supraglacial debris cover using a supervised classification approach on an ASTER image. Mapping the rock types present in a surface debris cover is valuable for 3 reasons. First, to identify spatial variation in emissivity, second, to identify spatial variation in thermal conductivity, and third, identification of the source areas of debris and transport pathways. Rock types present on the Miage glacier were mapped using a supervised classification approach, with the classified rock map showing that the rock type distribution could be mapped using ASTER data. These results were validated using field data, however, issues relating to the validation of map outputs with field data were highlighted. As the
large spatial resolution of the ASTER pixel (90 x 90 m) making comparison with point field data difficult, due to limitations in obtaining data in the field at a 90 x 90 m scale which corresponds to the location of an ASTER pixel.

Problems validating the rock map with field data were exacerbated by the mismatch in spatial scale of the ASTER classification output (90 x 90 m) and the high variability of rock types found on the Miage Glacier. This meant that small scale variability in rock types, which is particularly high in the lower reaches of the glacier, could not be identified as the variability is averaged out within the 90 x 90 m pixel. Consequently, only the dominant rock type in these pixels tends to be identified by the classification. A solution to this would be to apply a fuzzy/soft classification method which would identify the presence of other rock types within each pixel (e.g. Paasche and Eberle, 2009). Despite this limitation, the classified rock type distribution map demonstrates that the general patterns in rock type can be mapped using these data on the Miage Glacier.

9.3. Applicability of findings in both a glaciological and remote sensing perspective

9.3.1. Glaciological

When glaciers become covered in debris their dynamics and geomorphological processes are significantly affected due to the influence of the debris material on surface ablation rates and mass balance, resulting in a change of the behaviour from ‘clean’ glaciers (Popovnin and Rozova, 2002; Deline, 2005; Kirkbride, in press). However, the number of studies on debris-covered glaciers is significantly lower than those on ‘clean’ glaciers, which in turn limits the full understanding of how a debris layer on a glacier will modulate the amount of ice ablation in mountain regions under possible global warming impacts.
The discharge characteristics of meltwater streams emerging from glaciers are also affected due to the impact of surface debris upon ablation. The changing discharges that result from debris-covered glaciers must be monitored to predict the long-term availability of water resources as many are dependent upon this melt water as a freshwater resource (Kayastha et al. 2000; Stern, 2006). These include ~40% of agriculture in the Andean valleys (which in turn will affect ~50 million people). The Himalaya-Hindu Kush region would also be affected, where glacial meltwater feeds seven of Asia’s largest rivers, including the Ganges, with glacial melt contributing to 70% of summer flow and providing water to ~500 million people (Stern, 2006).

Therefore, methods which map debris thickness and extent, that are reasonably straight forward to apply, transferable in space and time, and importantly have realistic data requirements, have an important role to play in mass balance monitoring, sea level change prediction, global climate models, catchment scale hydrological models and glacial hazard monitoring. Especially, since at present global estimates of glacier response and contribution to sea level change do not incorporate melt runoff from debris-covered glaciers (Church et al. 2001; Kargel et al. 2005; Raper and Braithwaite, 2006).

An issue is whether key data inputs for these techniques are, or are likely to become available, at a global scale. In relation to DEM products, the spatial resolution of both SRTM and ASTER data are too coarse for many small glaciers, and have accuracy issues in areas of steep complex terrain, making their applicability limited. However, the development of higher spatial resolution regional DEMs is increasing, such as Switzerland’s 25 m DEM (Swisstopo, 2005), although this is not freely available outside Switzerland. Problems also occur in relation to the acquisition of suitable
visible and thermal imagery due to issues of cloud and/or snow cover which significantly limits the number of images available that can be used for analysis. However, with increasingly longer image archives, such as the Landsat (MSS, TM and ETM+) 37 year record, and ASTER 9 year record, the number of images which can be utilised are increasing. Although the potential of future monitoring is limited due to a lack of suitable sensors and platforms (section 8.3.2). Therefore, if these issues are met, then future work must concentrate upon the application and refinement of these mapping techniques, to ensure that debris extent and thickness changes can be measured on a global scale. Also, it will provide the data sets needed by both modellers and hazard managers to develop models to forecast how debris covers may change in the future.

The development of a method to estimate debris thickness and utilisation of DEMs to estimate elevation change from a satellite enables the identification of both debris thickness and its variation, and surface elevation changes over time at a much greater spatial scale than can be obtained by field based studies. Despite the spatial resolution of satellite images being less than that which could be obtained at the ground, the completion of a study at such a fine scale could not be completed in the field due to the size of these glaciers, making depth sampling, or ablation measuring (for elevation change studies) at a very fine spatial scale impossible. The temporal repeat frequency is also much greater than can be obtained with field based methods alone due to financial, access, and time issues.

9.3.2. Remote sensing

Through the development of new methods for monitoring debris-covered glaciers, and testing and refinement of others which have previously been applied, support is provided for the utilisation of remote sensing for debris-covered glacier monitoring. The suitability of the ASTER and Landsat
(MSS, TM, and ETM+) satellite sensors for this task has also been highlighted. However, the ASTER mission was identified as having only a 15 year lifespan when it was launched in 2000, although problems with its SWIR sensors have been occurring since April 2008 (ASTERweb, 2009) and currently has no planned replacements. Also, despite Landsat 5 (launched 1984) and 7 (launched 1999) still being operational, a number of technical problems have been encountered and both missions are currently overrunning their predicted lifecycles of at least 3 and 5 years, respectively.

Without continuation of the ASTER and Landsat (MSS, TM, and ETM+) satellites, or development of a new sensor with similar properties, the monitoring of debris-covered glaciers would be affected. This would limit future debris extent monitoring at a high temporal resolution, as aerial data could not be obtained for a large number of debris-covered glaciers at the same temporal resolution. Fortunately, a replacement Landsat satellite is being developed (Landsat Data Continuity Mission) and aims to be operational by 2012 (LDCM, 2007). However, launch has already been delayed from 2011 to December 2012 (LDCM, 2009) and other delays may still be encountered. Therefore, a gap in the image record may result if ASTER or the 2 Landsat sensors fail before the new Landsat sensor can be launched, which would have a significant impact upon both clean and debris-covered glacier monitoring at that time.

9.4. Future work

Throughout the completion of this study a number of limitations were encountered, including: problems in the calculation of $SHF$ values in the debris thickness model, use of $z_o$ as a tuning parameter in the debris thickness model, application of energy balance model to another glacier, estimation of the energy balance component of heat store in the debris thickness model,
implications of using an average emissivity value in surface temperature estimates from thermal imagery, and the source mapping of debris cover. These limitations have highlighted areas which require further investigation and should form the basis for future investigations. Investigation into these will increase the future applicability of the methods which were developed and tested.

First, when the model for debris cover thickness estimation was developed, some problems were encountered in the calculation of $SHF$ values at some pixels (which become significantly over-estimated) using a physically-based approach with extension to non-neutral conditions using the bulk Richardson number. This probably occurred due to incorrect air temperatures being estimated at some pixels, which generated a large temperature gradient in the surface layer. When combined with low windspeeds, this affected the $SHF$ calculations and resulting debris thickness estimates. Therefore, this highlighted a key problem of extrapolation of air temperatures on debris-covered glaciers. Because surface temperature and air temperature are closely related through $SHF$ and $LWR\downarrow$, the spatial pattern of surface temperature is more closely related to debris thickness and shortwave radiation receipts than elevation.

As a result, future work should focus on developing methods of retrieving air temperatures at each pixel using the relationship between air and surface temperature. However, a successful solution to the problem would also need to take into account other meteorological conditions influencing the surface energy balance. Therefore, it would need to be physically-based, rather than simply an empirical relationship to debris thickness. Given extensive existing datasets, including a map of debris thickness, the Miage Glacier would be an ideal site to develop it. From this, a more robust air temperature equation can be developed. This is, therefore, also important for distributed melt modelling.
The $z_o$ value in the energy balance model has also identified areas which require further investigation. In the energy balance model for estimating debris thickness, the $z_o$ value is used as a tuning parameter to ensure reliable debris thickness estimates are generated. However, due to the spatial variability of $z_o$ on a glacier’s surface, whether this would work as effectively at other sites needs investigation. Because the use of a single $z_o$ value may not produce accurate debris thickness estimates at all locations on a debris-covered glacier. It also highlights the requirement of some field data to ensure the $z_o$ value used is producing reliable debris estimates.

Another key area for future investigation is the testing of the energy balance model for debris thickness estimation on another glacier. The model has been successfully applied to another image of the Miage Glacier, but application to a different location is also important to demonstrate spatial transferability. This transferability has the potential to be a significant advantage over empirically based methods, which are known to have limited transferability (both in space and time). However, a number of issues may arise with the application to another debris-cover glacier which will need further investigation. First, the model needs to be tested on a glacier with a different slope and aspect to the Miage Glacier, as a fundamental simplifying assumption in the model is of a flat glacier, and this may not hold for a glacier with different topography. Any problems in debris thickness estimations with the energy balance model with this application at a different glacier will need to be resolved through the development of the model and input parameters used. Second, issues mentioned above and whether the $z_o$ value can be used as a tuning parameter on different debris-covered glaciers needs to be tested.
The energy balance component of heat store (and its variability) was only estimated as a constant percentage of \( COND \) in the model, as the rate of debris temperature change could not be calculated (as only one instantaneous satellite image was available). This estimated value, which identified that 36% of \( COND \) goes into heat store, was developed at the LWS and, therefore, does not consider any variation in heat store at different locations on the glacier. Further investigation into the component of heat store and its variability across the glacier is, therefore, required, as it will have an impact upon the debris thickness value estimated. Especially because the percentage of \( COND \) going into store is likely to decrease as debris cover gets thinner. To address this, the amount of \( COND \) which goes into heat store needs to be calculated at other locations on the glacier (such as the UWS) and any significant variability identified, and incorporated into the model.

Issues also arose when developing the model to include the \( SWR \downarrow \) and \( LWR \uparrow \) of each pixel, as a DEM with high spatial resolution from a date close to the image on which the energy balance method is being performed is required. However, the availability of such a DEM in many locations is limited, again highlighting that in those locations where data are not available, a more simplified model will again be the best solution. Also, field measurements (surveys) could be used to test a fully-physical model in small areas.

Chapter 8 outlined the development of an approach to map the rock types present on a debris surface, so that the emissivity of the different rock types present could be identified. The effect that different rock emissivity values have upon the retrieval of ASTER surface temperature estimates and resulting impact on debris thickness estimates using the energy balance method was also briefly investigated. Due to the limited published research in this area, future work needs to
focus upon the implications of using an average emissivity values to obtain surface temperature estimates from ASTER imagery. This could be investigated at the Miage Glacier, by identifying the variability in emissivity within a pixel and comparing the surface temperature estimate from the ASTER image (which uses an average emissivity value) to surface temperature values obtained using emissivity values found within the 90 x 90 m area.

It was also highlighted in chapter 8 that rock type classification maps could be used to identify the source of debris material. Therefore, debris cover movement could be inferred and the dominant source identified. Once identified, reasons as to why it is the dominant source can be identified, such as it is the most common rock in the area or whether it is more susceptible to erosion and/or rock wall destabilisation (through melting of ice). Also, once the location of these deposits is known, monitoring of the change in debris inputs over time can be completed, enabling the identification of any changes over time, with any changes in debris input sources possibly related to changes in climate. These include the thawing of rocks (permafrost degradation) further up the glacial valley sides in conjunction with climate warming, as this would alter (and increase) the debris source inputs onto a glacier surface. Consequently, the monitoring of these debris inputs on the valley sides would enable the identification of where and when permafrost degradation is occurring. Further analysis into the impact of emissivity variations dependent upon the rock types present upon sub-debris ablation rates is also required. This will enable improved predictions on meltwater from debris-covered glaciers, especially in response to future predicted warming global temperature trends.
9.5. Summary

Overall, this study has developed and tested a number of key applications for debris-covered glacier monitoring, including debris thickness mapping, debris extent monitoring, surface elevation and velocity measurement, and identification of rock types present on a debris layer. Increased research into these applications through addressing issues raised both throughout this project and in the future work section above will increase their applicability and ensure their transferability, enabling application at a global scale. This, in turn, will increase the amount of information available on debris-covered glaciers whilst minimising fieldwork requirements and, therefore, result in an increased understanding of both their behaviour and development, which is a key issue at present, as the number of debris-covered glaciers and extent of debris areas is likely to increase as a result of climate warming impacts (Diolaiuti et al. 2003; Paul et al. 2003; Kellerer-Pirklbauer et al. 2008).

This increase in the number of debris-covered glaciers will, in turn, have significant impacts upon populations which depend upon these glaciers as a freshwater resource. Also, it will have implications on the number of people at risk from debris-covered glacier hazards, since moraine dammed glacier lake outburst floods are a particular characteristic of retreating debris-covered glaciers (e.g. Saki et al., 2000; Benn et al., 2001; Quincy et al., 2007). The global monitoring of these debris-covered glaciers is, therefore, an essential process that will become increasingly possible (even on glaciers in remote locations) through the methods developed and tested during this project which can be incorporated into existing global monitoring schemes such as GLIMS. This will enable the collection of data and its analysis at a high temporal resolution on a large number of debris-covered glaciers. As a result, this study has contributed towards filling a void in both the data available on debris-covered glaciers and the methods to obtain this data.
CHAPTER 10: CONCLUSION

Previous research monitoring debris-covered glaciers has been sparse (especially when compared to work on ‘clean’ glaciers), with studies limited to boundary mapping (e.g. Taschner and Ranzi, 2002; Ranzi, et al., 2004), surface elevation changes and surface velocity measurements (e.g. Kaab, 2002; 2007; Luckman et al., 2007), identifying the rate of ablation under a debris layer (e.g. Nakawo and Young, 1981; Nakawo et al., 1993; Nicholson and Benn, 2006), and debris cover changes over time (e.g. Stokes et al., 2007; Bolch et al., 2008; Shukla et al., 2009). Such a limited focus is surprising considering the key impact a debris cover has upon a glacier, through its impact upon surface ablation rates, and resulting impact on mass balance studies and runoff estimates. All are of key importance in predictions of sea level rise and freshwater availability.

One of the main reasons for reduced monitoring of debris-covered glaciers is the difficulties in accessing these glaciers and problems of obtaining large amounts of data at a high temporal resolution. Remote sensing has been widely applied to ‘clean’ glacial monitoring and a number of techniques are currently used in global glacier monitoring programs, including GLIMS. However, remote sensing has had limited application to debris-covered glaciers despite its potential for increasing the amount of data available on these glaciers. Developing remote sensing as an effective tool for monitoring will enable increased understanding of debris-covered glacier systems, which is a key requirement for understanding the Earth system since the number of debris-covered glaciers is expected to increase due to the impacts of global climatic warming.

Consequently, this study focussed upon the development and implementation of new techniques for debris-covered glacier monitoring, and the testing of previously developed algorithms to
assess their effectiveness at different sites. The development and testing of these methods in this study aimed to highlight their applicability for debris-covered glacier monitoring, and provide methods which could be incorporated into existing global monitoring schemes, such as GLIMS. The incorporation into global monitoring schemes would ensure that increased monitoring of debris-covered glaciers can be completed. The main findings of this study which highlight the successes and failures of these methods are included below.

- The ability to map debris thickness and its variability over time is of key importance in mass balance monitoring and investigation of surface elevation changes over time. This study successfully developed and tested a physically-based energy balance model to estimate patterns of debris thickness on the Miage Glacier, requiring basic meteorological variables recorded at the site, and surface temperature from an ASTER AST08 image. Results showed good general agreement with field measurements. Although similar results can be obtained from empirical relationships of debris thickness to surface temperature, the energy balance approach is a better alternative since it has the potential to map debris covers over wide areas with only limited field data. The energy balance model also has the advantage of transferability across space and time compared with empirical relationships, which are dependant on intensive fieldwork to provide site specific calibration data, and hence incorporate a degree of stationarity.

- Mapping of supraglacial debris-cover extent and its change over time is an important component in the monitoring and modelling of the response of glaciers to climate change. The application of two different methods during this study has shown the potential of satellite-based methods (manual digitising, and a semi-automatic approach) to map debris cover extent on the Miage Glacier. It also successfully demonstrated the application of the semi-automatic approach of Paul et al., (2004) to a different satellite sensor (ASTER), which is important if such methods are to be used at different sites and with different image data in the future.
• Monitoring of surface elevation changes over time provides a way of monitoring the impact of climate changes upon debris-covered glaciers, since the frontal recession of a debris-covered glacier is a poor indicator. This is a consequence of the main cause of mass loss being downwasting, since the thick debris cover at the snout prevents any retreat. However, the application of ASTER DEM data at the Miage Glacier failed to accurately determine surface elevation changes between 2000-2006. The reason for this was the problems of ASTER DEM accuracy in regions of steep and varied terrain, which was highlighted as very poor in this study with errors of 1000’s of metres on some DEMs. Therefore, the potential for ASTER DEMs for this application was limited, as other studies in other locations have also encountered similar problems (e.g. Kaab 2002, Bolch et al., 2008). In turn, the need for the improvement of ASTER DEMs is identified, if this cannot be achieved the requirement for DEMs with increased accuracy and high temporal resolution like ASTER in these regions is demonstrated.

• Surface velocity variations also provide a means of identifying the impact of climate changes on a debris-covered glacier, since velocities will slow when ice inputs are reduced, resulting in the stagnation of a glacier. ASTER DEM data was again used to estimate velocities, however, this analysis encountered similar problems to the surface elevation change analysis, with DEM errors causing problems with surface velocity estimation. An example of this was the identification of movement of the mountain valley sides which is clearly impossible. One reason for this may have been the application of a DEM with some cloud and snow present which is known to cause ASTER DEM errors, therefore, this requires further testing on a better quality ASTER DEM. However, the availability of ASTER DEM data with no snow or cloud in these regions is difficult and once again highlights the limitation of using ASTER DEM data in this region, on a glacier bounded by steep valley sides.
Different rock types have different emissivity and albedo values, this leads to differential ablation depending upon the rock types present causing spatially variable ablation rates. It also has an impact upon the retrieval of surface temperature in remotely sensed imagery which utilises an emissivity value. ASTER imagery was utilised for this application, and was able to distinguish the presence of different rock types present on the glaciers surface through a supervised classification approach. The application of an unsupervised approach also highlighted the potential of this method, even when limited or no field data are available. In turn, this demonstrates the potential of ASTER imagery to identify the variability in emissivity on a debris-covered glacier. In turn, variable ablation levels can be identified and incorporated into mass balance studies and calculated melt rates from debris-covered glaciers. Knowing the emissivity values present within an area will also enable a more accurate determination of surface temperature in satellite imagery, which in turn will improve debris thickness estimations.

10.1. Concluding Comments

Overall, the potential of ASTER data in the monitoring of debris-covered glaciers and debris thickness estimations, debris extent variations, and mapping of different rock types present on a glacier has been demonstrated. However, problems with the ASTER DEMs have identified areas of concern for their application for both surface elevation changes and velocity variations over time. This highlights the requirement for a globally available DEM at a high temporal resolution with improved accuracies in areas of steep and varied topography. Whilst remote sensing remains an attractive opportunity to monitor debris-covered glaciers, there remains significant uncertainty in both data and understanding.


Deline, P. & Orombelli, G. (2005) Glacier fluctuations in the Western Alps during the Neoglacial, as indicated by the Miage moranic amphitheatre (Mont Blanc massif, Italy), *Boreas*, 34, 456-467


Earth remote sensing data analysis centre (ERSDAC) (1996) Algorithm theoretical basis document for ASTER level-1 data processing (version 3.0), ERSDAC

Earth remote sensing data analysis centre (ERSDAC) (2001) ASTER users guide part 1 - general (version 3.1.), ERSDAC

Earth remote sensing data analysis centre (ERSDAC) (2003) ASTER users guide part 2 - level 1 data products, ERSDAC

Earth remote sensing data analysis centre (ERSDAC) (2005) ASTER users guide part 3 -DEM, (Version 1.1), ERSDAC

Earth remote sensing data analysis centre (ERSDAC) (2005) ASTER users guide part 3- Ortho, (Version 1.1.) ERSDAC

Etzelmuller, B. (2000) Research article on the quantification of surface changes using grid-based digital elevation models (DEMs), Transactions in GIS, 4, 129-143


Kaab, A. (2005) Remote sensing of mountain glaciers and permafrost creep, Physical Geography Series, 48, University of Zurich


Konrad, S. K. (1998) Possible outburst floods from debris-covered glaciers in the Sierra Nevada, California, Geografiska Annaler, Series A Physical geography, 80, 3-4


Landsat, Global Orthorectified Landsat Datasets, http://www.landsat.org/dataservices/Landsat_ortho/Dataset_Description.htm (accessed 25/05/08)


Moghtaderi, A. Moore, F. Mohammadzadeh, A. (2007) The application of advanced space-borne thermal emission and reflection (ASTER) radiometer data in the detection of alteration in the Chadormalu paleocrater, Bafq region, Central Iran, Journal of Asian Earth Sciences, 30, 238-252


Munro, S. D. (1990) Comparison of melt energy computations and ablatometer measurements on melting ice and snow, Arctic and Alpine Research, 22, 153-162


Nakawo, M. & Young, G. J. (1981) Field experiments to determine the effect of a debris layer on ablation of glacier ice, Annals of Glaciology, 2, 85-91


Ostrem, G. (1959) Ice melting under a thin layer of moraine, and the existence of ice cores in moraine ridges, Geografiska Annaler, Series A Physical Geography, 41, 228-230


Swinbank, W. C (1963) Long-wave radiation from clear skies, Quarterly Journal of the Royal Meteorological Society, 89, 339-348

Swisstopo (2005) DHM25 the digital height model of Switzerland, Federal Office of Topography


Tinti, S. Maramai, A. Cerutti, A. V. (1999) The Miage Glacier in the valley of Aosta (Western Alps, Italy) and the extraordinary detachment which occurred on August 9, 1996, Physics and Chemistry of the Earth, 24, 157-161


APPENDIX – Published works relating to this thesis

Published in the conference proceedings of: International Symposium on Remote Sensing of the Environment, May 4-8 2009
Abstract – Debris-covered glaciers (>50% of the ablation zone covered by debris) are common in many of the world’s major mountain ranges and represent an important freshwater resource. Previous studies have established that surface melt rates decrease strongly for an increase in supraglacial debris thickness from a few cm up to 20-30 cm, whereas, thin and patchy debris cover enhances melt due to albedo reduction. It is therefore important to develop methods to map and monitor changes in debris cover to enable predictions of glacier responses to climate change through numerical modelling. It is not possible to achieve this from ground-based measurements alone. Therefore, a method to estimate debris thickness from an ASTER thermal band surface temperature image, based on a physical solution of the energy balance at the debris surface, has been developed. The model performs well in comparison to previous empirical methods showing good agreement with measured debris thicknesses.

Keywords: ASTER, Debris covered glacier, surface energy balance, glacier melt model

1. INTRODUCTION

Supraglacial debris thickness and surface temperature are two key variables required to calculate sub-debris ablation rates on debris-covered glaciers (Mihalcea et al. 2008). Field measurements are time-consuming and difficult in many locations, highlighting the potential of remote sensing to resolve the problem. Development of a model to calculate sub-debris ablation using remote sensing would be a valuable tool, particularly in inaccessible locations and over extensive areas. However, there have been few previous attempts to estimate debris thickness from satellite imagery despite the potential advantages, including rapid data acquisition, ability to access difficult to reach locations and the possibility to estimate debris thicknesses at higher spatial and temporal frequencies than ground measurements (Ranzi, et al., 2004).

A simple model that will calculate supraglacial debris thickness on a glacier has been developed. This model combines TERRA ASTER thermal imagery (to provide surface temperature) with measured meteorological variables including surface radiation fluxes and extrapolated air temperature to solve the surface energy balance for each pixel. Based on assumptions about the thermal properties of the debris, debris thickness, \(d\), is found each pixel as a residual. As the method is physically-based it is potentially transferable in time (i.e. to monitor changes at a site) and in space to other glaciers. In this paper, results from the physical model are compared with a previous empirical (i.e. non-transferable) method of calculating \(d\), which was calibrated for the study site and same ASTER image using field measurements of debris thickness and surface temperature (Mihalcea et al. 2008).

2. STUDY SITE

Miage glacier (45°47’30”N, 6°52’00”E) is a temperate glacier situated on the Italian side of the Mount Blanc Massif in the western Alps of Italy, a few kilometres west of Courmayeur in the Valley of Aosta. It is the largest (11-13km) and most representative debris-covered glacier in the Italian Alps (Pelfini, et al. 2007), and is morphologically similar to large Asian debris-covered glaciers (Smiraglia et al. 2000). The glacier is extensively debris-mantled from the snout at 1770 m above sea level (a.s.l.) up to an altitude of approximately 2400 m a.s.l., covering over 6 km of the glacier tongue in total (Deline and Orombelli 2005). Above this elevation debris cover is discontinuous except on two large medial moraines. Debris is supplied mainly by rockfalls and avalanche events at exposed headwalls between the tributary glaciers (Thomson et al. 2000, Pelfini, et al. 2007).

3. MODEL DEVELOPMENT

To determine debris thickness from ASTER thermal imagery (AST08) a model was developed based on the surface energy balance equation (e.g. Brock and Arnold 2000, Equation 1). A bulk aerodynamic profile method, using the Richardson number to account for variations in atmospheric stability, is used to calculate the sensible heat flux (SHF). Energy balance components are calculated using measurements of net shortwave radiation (SWR), net longwave radiation (LWR), air temperature \(T_a\), and wind speed \(u\) at 2 m (initially all assumed constant across the glacier) at an automatic weather station (AWS) located on the glacier at 2030 m a.s.l. Surface temperature \(T_s\) was provided by an ASTER surface temperature image for 01/08/05 at 10:40 UTC. Standard values of atmospheric constants were used with an initial aerodynamic roughness length \(z_0\) of 0.001 m. As the surface was dry at the time of image acquisition, the latent heat flux, \(LHF\) can be ignored. Similarly, as the rate of debris temperature change is unknown when only one satellite image is available, the flux of change in stored heat energy, \(\partial STOR\), cannot be calculated, resulting in a simplified energy balance (Equation 2). Equation 2 is solved for every pixel to find the conductive heat flux (COND) as a residual. By assuming that the ice-debris interface is at melting point and that thermal conductivity \(K\) is uniform across the glacier, debris thickness \(d\) is calculated from a re-arranged heat conduction Equation 3.

\[
\begin{align*}
SWR + LWR + SHF + LHF + COND + \partial STOR &= 0 \quad (1) \\
SWR + LWR + SHF &= COND \quad (2) \\
K \times T_s/COND &= d \quad (3)
\end{align*}
\]

Where: \(K = 0.96 \text{ (Wm}^{-1}\text{K}^{-1})\), the average value of debris thermal conductivity calculated at 25 point locations (ablation stakes) on Miage glacier (Brock et al., submitted).
Ignoring $\partial_{STOR}$ leads to an overestimate of $COND$, as the debris is warming at the time of ASTER image acquisition. On average, 36% of the $COND$ flux (Equation 2) goes into heat store between 10:00 and 11:00 am under fine conditions, based on calculations over the whole 2005 ablation season using measured $T_S$. The $COND$ flux was therefore reduced by 36% before calculating $d$.

### 3.1 Point application of the model to the AWS site and air temperature extrapolation

To test the validity of the simplified energy balance model, $d$ was calculated each hour on 01/08/05 using measured meteorological variables, and $T_S$ values calculated from the measured upwelling longwave radiation flux. The calculated value of $d$ remains reasonably constant during the period of the day when the assumption of a linear downward temperature profile in the debris is most likely to be met (Figure 1). This lends support to the application of the model at a glacier-wide scale.

![Figure 1: Estimated $d$ from the energy balance model at the AWS site on 01/08/05 ($z_o = 0.001$ m) for the time period when the assumption of a linear downward temperature profile in the debris is likely to be met. Point $d$ measured at the AWS site is 0.16 m.](image)

The model calculated $d$ at the AWS site at 10:40 on 01/08/05 is 0.09 m for $z_o = 0.001$ m, increasing to 0.15 m for $z_o = 0.01$ m. This is close to the measured $d$ of 0.16 m (although this point measurement does not necessarily correspond to the 90 x 90 m area sampled by the ASTER pixel). The $z_o$ value used in the model clearly has a significant impact upon the resulting $d$ value. Little is known about the magnitude of $z_o$ on debris covered glaciers, however, and hence it will be treated as a tuning parameter in this study.

When the model was applied to the rest of the glacier, assuming a constant $T_{air}$, incorrect estimates of $d$ resulted, even for locations at similar elevation to the AWS. This problem arises because $T_{air}$ does not conform to a standard lapse rate on a debris-covered glacier during the daytime, but is instead closely related to $T_S$ through surface convection. Hence, $T_{air}$ must be calculated from values of $T_S$. Available data are insufficient for a physically-based solution at all sites, so instead an empirical relationship between $T_{air}$ and $T_S$ was developed based on regression of values recorded at the AWS on 01/08/05, where $T_S$ was derived from measurements of the upwelling longwave radiative flux.

Regression relationships were developed for different time periods, with the relationship using data between 08:00-10:00 (Equation 4), giving the best $T_{air}$ estimates (Figure 1). This is probably because heat transfer between the debris and surface atmospheric layer through convection is very strong in the mid morning. A regression equation was also developed using daytime data (08:00 – 20:00) from the entire summer period in 2005 (85 days), to provide a more widely applicable relationship (Equation 5). However, none of the equations account for the highly variable nature of $T_S$ over a 24 hour period (Figure 2).

\[
\text{2 m air temperature} = 5.79 + 0.35 \times T_S \quad (4)
\]

\[
\text{2 m air temperature} = 3.36 + 0.46 \times T_S \quad (5)
\]

![Figure 2: Plot showing comparison of $T_{air}$ estimates calculated using different regression equations generated using measured $T_{air}$ and $T_S$, alongside plot of measured $T_{air}$ on 01/08/05](image)

### 3.2 Model application to the whole glacier area

Once $T_{air}$ had been extrapolated to each ASTER pixel using Equation 4, $d$ was estimated for all pixels on the glacier, using $z_o$ values in the 0.001 – 0.01 m range. $z_o$ values of <0.002 m and >0.005 underestimated $d$, respectively.

### 4. MODEL VALIDATION

Estimated debris thicknesses from the energy balance model are plotted in Figure 5a and b. The results are compared (visually and numerically) to $d$ values from the empirical method of Mihalcea et al. (2008) and field measurements of $d$.

#### 4.1 Measured thicknesses

Estimated $d$ values from the energy balance model were compared to measured debris thickness collected during fieldwork in 2005 (single point locations), 2006 (single point locations and transects), and 2007 (transects). Transects provided the best comparison to the ASTER $d$ as an average.
value from a 90x90 m transect (with debris thickness measured at 10 m intervals on four transect arms) provides a value on the same spatial scale as the ASTER pixels, with small scale variations less prominent (Table 1). The RMSE for the 25 point sites is 0.16 m for $z_0 = 0.003$ m and 0.09 m for $z_0 = 0.002$ m. The bigger RMSE for $z_0 = 0.003$ m is mainly due to a large overestimation at just two sites (5 and 6, Figure 3). The results from the transects show that, excluding the upper reaches site in 2007, and middle reaches site in 2006 estimated $d$ is within a few cm of the measured $d$ (Table 1). It must be borne in mind, however, that even in the transect measurements, $d$ is not sampled over large areas of an ASTER pixel, so exact agreement between modelled and measured values would not be expected.

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<tr>
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<th>Measured thickness (m) (Transect average)</th>
<th>Debris estimate $z_0 = 0.002$ (m)</th>
<th>Debris estimate $z_0 = 0.003$ (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>2006 lower (100 m² area)</td>
<td>0.23</td>
<td>0.17</td>
<td>0.19</td>
</tr>
<tr>
<td>2006 middle (100m² area)</td>
<td>0.18</td>
<td>0.09</td>
<td>0.09</td>
</tr>
<tr>
<td>2006 upper (100 m line)</td>
<td>0.17</td>
<td>0.22</td>
<td>0.27</td>
</tr>
<tr>
<td>2007 Upper (area 45x30 m)</td>
<td>0.17</td>
<td>0.06</td>
<td>0.06</td>
</tr>
<tr>
<td>2007 Lower (100m² area)</td>
<td>0.20</td>
<td>0.17</td>
<td>0.20</td>
</tr>
</tbody>
</table>

Table 1: Comparison of measured (in the field) and estimated $d$ (from the model with two different $z_0$ values used)

4.2 Comparison with previous studies

Results from the physically-based model (Figure 4 and b) show broad agreement with a map of field-measured debris thicknesses collected in 1997 (Figure 4d). However, movement of the debris in the intervening 8 years since 1997 due to ice flow invalidates a detailed comparison. Encouragingly, the model replicates well the known broad patterns of the debris thickness distribution on Miage glacier. In particular, the thick debris on the two terminal lobes and on the medial moraines on the lower valley section of the tongue are well reproduced, as is the thin debris on the upper tongue and in heavily crevassed areas just above the divergence of the terminal lobes. Some anomalies are also apparent, such as the prediction of thin debris on the southern edge of the northern terminal lobe. Possibly, incorporation of some extra-glacial vegetated ground in these pixels could account for the low $T_s$ and $d$ values here.
The broad patterns of $d$ resulting from the empirical method of (Mihalcea et al. 2008) are similar (Figure 4c). The empirical method identifies more variation in thickness on the upper tongue of the glacier, but the energy balance method is better at identifying areas of greater thickness ($d = 0.4 - 0.5$ m) on the medial moraines in the middle reaches of the glacier and thicknesses greater than 0.6 m at the snout. Despite these differences, the thicknesses estimated on the lower sections of the glacier are very similar in both approaches, with the thickest debris close to the terminus contrasting with patches of debris <0.20 m above the divergence of the terminal lobes.

5. CONCLUSION

This paper has described the development and testing of a physically-based energy balance method to estimate patterns of debris-cover thickness on a glacier, which requires basic meteorological variables recorded at the site and $T_s$ data from an ASTER image as input. Initial results using an image of Miage glacier acquired at 10:40 UTC on 01/08/05 are promising, showing good general agreement with field measurements. While similar results can be obtained from empirical relationships of $d$ to $T_s$ (e.g. Mihalcea et al. 2008), the energy balance approach is a better alternative since it has the potential to map debris covers over wide areas with only limited field data. The energy balance method also has the advantage of transferability in space and time compared with empirical relationships which are dependent on intensive fieldwork to provide site-specific calibration data. The main issues to be addressed in future work are: i) a need for improved methods to extrapolate air temperature across debris-covered glaciers; ii) inclusion of the flux of change in heat store in the debris layer; and iii) improving understanding of spatial variations in $z_0$ and debris properties such as $K$. Further improvements to the model will be expected with future developments to incorporate local slope and aspect in individual pixels, using a high resolution DEM, and to vary the net longwave radiation flux spatially, by estimating the upwelling flux based on the ASTER-derived surface temperature.

REFERENCES


Deline, P. & Orombelli, G. “Glacier fluctuations in the Western Alps during the Neoglacial, as indicated by the Miage moranic amphitheatre (Mont Blanc massif, Italy)” Boreas, vol 34, p. 456-467, November 2005.


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Monitoring changes in debris-cover extent on Miage glacier, Italian Alps, between 1975-2006 using optical and thermal satellite imagery

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Abstract - Identification of debris-cover extent and its change over time on glaciers is an important challenge within the framework of GLIMS (Global Land Ice Measurements from Space) project. Methods based on multispectral analyses are hampered due to the spectral characteristics of supraglacial debris, which are similar to surrounding non-glacial terrain. To resolve this, previous studies have applied manual approaches through on-screen digitising of debris-covered areas. Despite promising results, this method is labour intensive and time consuming, especially for a large number of glaciers. Consequently, semi-automatic methods have been developed. This study tests the applicability of both manual digitizing and a semi-automatic approach (Paul et al., 2004) to the debris-covered Miage glacier. Results are positive but show only a small increase in the areas of continuous debris cover between 1990-2004, implying the transition to extensive debris-cover on this glacier occurred prior to 1990.

Keywords: Debris covered glacier, Debris extent monitoring, ASTER, Landsat

1. INTRODUCTION

The monitoring of debris-cover extent variations is important as it has a direct impact on glacier mass balance: continuous debris more than a few centimetres thickness reduces ablation, whereas thin and patchy debris increases melting through albedo reduction (Brock et al. 2000). Furthermore, an increase in debris extent over time may indicate a glacier is in negative mass balance and vice versa (Deline, 2005). Debris cover on the Miage has varied significantly over the Holocene ranging from complete cover on the glacier tongue during warm periods, to an almost debris-free state in cold periods of glacier advance, most recently in 1770 (Deline 2005). Debris cover on the Miage has varied significantly over the Holocene ranging from complete cover on the glacier tongue during warm periods, to an almost debris-free state in cold periods of glacier advance, most recently in 1770 (Deline 2005). Several recent glacial studies (including: Kirkbride and Warren, 1999, Stokes et al. 2007, Kellerer-pirklbauer et al. 2008, Bolch et al. 2008) have reported increases in debris-cover extent in the late 20th century. This has been related to strongly negative mass balances and high summer temperatures which have both increased rates of debris supply, through permafrost thawing and destabilization of surrounding rock slopes (Kirkbride and Warren 1999, Stokes et al. 2007), exhumation of englacial debris bands due to high ablation rates and reduced glacier transport rates due to lower ice flux (Kirkbride and Warren 1999).

2. STUDY SITE

The Miage glacier (45° 47’ 30” N, 6° 52’ 00”E) is situated on the Italian side of the Mont Blanc Massif in the western Alps. It is the longest (11 km) debris-covered glacier in the Italian Alps (Pelfini, et al. 2007), and is morphologically similar to large high-Asian debris-covered glaciers (Smiraglia et al. 2000). The glacier is continuously covered in rock debris from the snout at 1770 m above sea level (a.s.l.) to around 2400 m a.s.l., above which debris cover is thin and patchy, except on medial moraines. Overall debris covers approximately 6 km of the glacier tongue (Deline and Orombelli 2005). Debris is supplied mainly through rockfalls and avalanches at exposed headwalls between the tributary glaciers (Pelfini, et al. 2007).

3. METHODOLOGY

Debris-cover extent variations were mapped using a number of images from different dates. The images used in this study consisted of three Landsat scenes: (13/07/75 (MSS), 10/09/90 (TM), 25/07/99 (ETM)), and four ASTER scenes (02/07/00, 14/08/04, 01/08/05, and 26/06/06). Two debris-mapping methods were tested. Firstly a manual approach based on visual assessment of debris-covered areas utilising on-screen digitizing, and secondly, a semi-automatic approach based on that of Paul et al. (2004), which combines multispectral satellite imagery with a digital elevation model (DEM) which was initially developed on Oberaletschglletscher in the Swiss Alps.

3.1. Manual approach

A particular challenge in the manual digitising was discriminating the upper ice/debris boundary, due to the problem of mixed pixels. Another problem became apparent in that areas on the glacier (and its tributaries) known to be debris free (from field observations) were found to have been misclassified as debris. These areas appeared darker than the bare ice areas (Figure 1a and b) due to a covering of fine rock dust on the surface (which lowered albedo) (Brock et al. 2000) and appeared visually similar to debris in the imagery.

Figure 1: Field photographs showing dust-darkened bare ice above the upper debris limit, a) tributary Glacier de Mont Blanc b) view upglacier towards the accumulation zone, June 2007

Manual delineation of debris cover extent may therefore lead to misclassification of bare ice as debris cover, especially on images following periods without rainfall when dust may accumulate on bare ice surfaces. In addition, all debris had to be classified as one type, as sporadic debris-cover cannot be discriminated from the continuous debris at the relatively coarse spatial resolution of the Landsat and ASTER imagery.
The ability to discriminate areas of sporadic debris would be very useful for distributed melt and mass balance modelling, however, as these areas experienced enhanced ablation, due to albedo darkening, in contrast to areas of continuous debris (Brock et al. 2000). This capability would require imagery of 5 m or finer spatial resolution, however.

The precision of the manual method was tested. Following Stokes et al. (2007), the precision error was calculated by digitising the debris outline 20 times and recording the maximum distance between any two digitised lines. True accuracy can only be assessed through comparison with field-based mapping, however. Delineation is commonly performed by one person to maintain consistency (Bolch et al. 2008). An alternative approach would be for more than one operator to digitise the same areas, and to target parts where outlines differ for more careful delineation.

3.2. Semi automatic approach

Automated classification may be able to overcome the problem of subjectivity apparent in the manual method, although mixed pixels, which are present in all satellite imagery (including modern fine spatial resolution sensors) due to the high spatial variability of debris cover, may still lead to mis-classification. Furthermore, multispectral classification through processes such as band ratios alone may not be able to separate debris-covered ice from extra-glacial debris (Paul et al. 2004), necessitating the use of additional information, such as a digital elevation model (DEM).

The method of Paul et al. (2004) was selected, as other semi-automatic methods, e.g. Taschner and Ranzi (2002) and Ranzi et al. (2004), were developed to identify the boundaries of debris-covered glaciers, rather than debris extent on the glacier surface, and testing of these methods on the Miage glacier (at the start of this study) proved their unsuitability for debris extent identification. We also tested the transferability of the Paul et al. method, initially developed for Landsat imagery, to the ASTER sensor, on the assumption that similar red, near infrared, and middle infrared channels on each sensor would produce comparable results (Paul et al. 2004). The only difference is the band ratio used, with TM4/TM5 used on Landsat images and VNIR3/SWIR4 on ASTER. Additionally, a high (10 m) resolution DEM, processed from stereo ortho-rectified aerial photographs, was used.

The method involves a sequence of steps, each excluding areas on the image which are determined not to be debris-covered ice, resulting in a final image showing only debris-covered ice areas:

1) The production of a band ratio image (Figure 2a), to reduce the impacts of variations in topography such as shadow.

2) The ratio image is then divided into two classes of ‘glacier’ (black) and ‘other’ (white) areas, classified using a threshold value of 2.0 (Figure 2b). The threshold was verified by visual inspection, as areas of glacier in shadow are sensitive to this threshold value. The lower the threshold the greater the number of partly debris covered pixels that are included, but at the expense of more noise elsewhere in the image (Paul et al. 2004).

3) An Intensity Hue Saturation image (IHS) is produced using bands TM3, 4, and 5 or ASTER bands 3, 4 and 5 (Figure 2c). This image enhances the three channels in relation to their colour contrast to separate the debris from other features on the image.

4) The Hue component of the IHS image was used to map vegetation (black) and vegetation free areas (white) using a threshold of 126 (Figure 2d).

5) Using a DEM all slopes >24° are excluded (Figure 2e). The value of 24° was based on the assumption that debris rarely rests on slopes steeper than this threshold value. This slope is also not exceeded for most debris covered tongues. This means that most vegetation-free and ice-free, mountain slopes can be excluded.

6) Each of the maps are overlain, masking those areas which are not debris covered so that only debris covered ice areas are shown in white (Figure 2f). Ideally, a glacier mask should be applied so that extra-glacial areas classified as debris-covered ice are not included.

Figure 2: Semi-automatic processing steps on ASTER 14/08/04 image: a) VNIR3/SWIR4 band ratio (glacier areas white), b) VNIR3/SWIR4 ratio image with threshold of 2 applied, (glacier areas black), c) IHS image bands 3, 4 and 5, d) IHS channel 2 with threshold of 126 applied, therefore, all vegetation removed (glacier areas black), e) Slope image showing areas <24° (white), f) debris extent (white)

4. RESULTS

The debris extent map generated from the semi-automatic method (Figure 2f) is a visually good approximation of the debris extent on the Miage glacier, although a number of linear areas on the tongue are misclassified as debris-free ice (shown as black). These features are in reality the debris-covered side slopes of large medial moraines on the tongue of Miage glacier. These were misclassified as being debris-free due to their slope
angle $>24^\circ$ (Figures 3a and 3b). The threshold slope angle in Step 5 was experimentally increased to see if these debris-covered slopes could be correctly classified. While it was possible to achieve this aim by increasing the threshold angle to $45^\circ$, the downside was the amount of white ‘speckle’, misclassified as debris-covered ice on the steep mountainous terrain and valley bottom increased. Thus, some manual correction of the final map is needed. Distinctive linear features, completely surrounded by debris-covered areas, on a glacier tongue should be easy to identify in such a process.

Unfortunately, the orthophoto DEM only covered the main tongue of the glacier below about 2500 m a.s.l., meaning that areas above the confluence of the main tributary glaciers could not be analysed. To resolve this an ASTER-derived DEM, covering the entire glacier, was applied. However, it was found that the fine detail picked out by the orthophoto DEM was lost and outputs were much more coarse due its poorer spatial resolution (30 m). The use of a high spatial resolution DEM is therefore advantageous. However, where this is not available, ASTER DEM products do provide a satisfactory input to automated debris-extent mapping on Alpine glacier, and may indeed perform better on the larger debris covered glaciers in the Himalaya or Andes (Paul et al. 2004).

Both methods were applied to all images (excluding the 1975 Landsat MSS image where the semi-automated method could not be applied due to its lack of middle infrared bands). However, several of the images were found to be badly affected (for our purposes) by snow and cloud cover, making the identification of debris extent difficult or impossible, particularly at the upper boundary of the debris cover. This illustrates the difficulty of satellite-based mapping of debris extent in mountain areas frequently affected by cloud and snow. Only two images proved suitable for assessing change in debris extent over time (the 10/09/90 Landsat TM image and 14/08/04 ASTER Image.)

Our results show that only a small increase in debris-cover extent occurred between 1990-2004 at the upper debris limit on the west side of the tongue (Figure 4 and Table 1). This increase is clearest in the manual method but is also evident to a lesser extent in the semi-automatic method (Figure 4). This confirms previous findings (Deline, 2005) that the transition to complete debris cover across most of the ablation zone of Miage glacier occurred prior to 1990. The current negative mass balance status of the glacier (Deline 2005) is not therefore being expressed through a marked increase in debris cover. This contrasts with studies on other debris-covered glaciers, which have reported an increase in debris-cover extent since 1990, which is thought to be associated with strongly negative mass balances and high summer temperatures in recent years (Kirkbride and Warren 1999, Stokes et al. 2007).

The similarity of the debris-extent maps from the two methods (Figure 4), supports the transferability of the semi-automatic method of Paul et al. (2004) to both other locations and other optical satellite sensors (ASTER). Some differences in extra-glacial areas are evident, however, in particular the threshold of 126 in the IHS image appears to be less effective in removing all vegetation in Landsat imagery (Fig. 4aii) than in ASTER imagery (Figure 4bii).

<table>
<thead>
<tr>
<th>Year</th>
<th>Debris Area - Manual method (km$^2$)</th>
<th>Debris Area - Semi-automatic method (km$^2$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1990</td>
<td>4.15</td>
<td>3.88</td>
</tr>
<tr>
<td>2004</td>
<td>4.25</td>
<td>4.064</td>
</tr>
</tbody>
</table>

Table 1: Debris covered area in 1990 and 2004

<table>
<thead>
<tr>
<th>Year</th>
<th>Max. difference (m)</th>
<th>Min. difference (m)</th>
<th>Average difference (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1990</td>
<td>160 (~5 pixels)</td>
<td>24 (~1 pixel)</td>
<td>46 (~2)</td>
</tr>
<tr>
<td>2004</td>
<td>107 (~7 pixels)</td>
<td>21(~1 pixel)</td>
<td>49 (~3)</td>
</tr>
</tbody>
</table>

Table 2: Precision results showing largest, smallest and average difference between 20 digitized lines of debris extent, sampled at 20 points

The precision analysis suggests a positional uncertainty of less than 2 or 3 pixels on average (Table 2) in the manual digitising method, with greater errors occurring at debris/ice boundaries where mixed pixels are present.

5. CONCLUSION

The mapping of supraglacial debris-cover extent and its change over time is an important component in the monitoring and modelling of the response of glaciers to climate change. This
paper has examined the potential of two satellite-based methods to map debris-cover extent, on Miage glacier: manual digitising based on visual assessment of the glacier surface, and a semi-automatic approach (Paul et al., 2004) which combines visible and near-infra-red bands with surface slope values to identify debris-covered areas of a glacier. The former approach is time-consuming, labour intensive, and potentially subjective, while the latter requires a high quality-high resolution (25 m or better) DEM and has the potential for misclassification, particularly if glacier limits are unknown. Our results support the application of both methods at this site, as similar results were obtained. Furthermore, the Paul et al. (2004) semi-automatic method is found to be transferable to different optical satellite sensors with similar wavebands to Landsat. Only a slight increase in the areas of debris-cover on Miage glacier was identified between 1990 and 2004, highlighting that the most of the expansion of debris-cover across the glacier tongue occurred prior to 1990.

A particular problem was the obscuring of the target by extensive cloud and/or snow which hampered full delineation of debris cover in several images. The acquisition of cloud- and snow-free images is problematic in temperate glaciated mountain ranges and the long repeat time-scale of applicable sensors (ASTER 48 days, Landsat MSS 18 days, TM 16 days, ETM 16 days) limiting the number of available images per year (although some sensors can be tasked to reduce the revisit time). The likelihood of finding a greater number of better quality images (cloud and snow free) increases as the timescale of study increases. Therefore, there is greater potential for identifying debris-extent variations over longer time periods than short, e.g. inter-annual, variations.

6. RECOMMENDATIONS

Several recommendations can be made for satellite-based mapping of supraglacial debris extent, from the findings of our study:

1) Whenever possible satellite imagery from the late ablation season should be used to minimise snow cover. A drawback in this, however, is that darkening of bare-ice areas by dust can be greatest at this time, potentially creating problems for manual discrimination of debris-covered areas.

2) Cloud-free scenes should be used, although care must be taken in selecting images due to the tendency for cloud-free conditions to occur after the passage of a cold front, which frequently deposits snow in the late ablation season.

3) High spatial resolution imagery provides more information at a finer spatial scale – enabling identification of boundaries more easily and reducing problems of mixed pixels, and is necessary, in particular, for small glaciers, where 1 m resolution imagery may be needed.

4) A high quality-high spatial resolution (at least 25 m) DEM is needed to apply the semi-automatic approach. Lower resolution DEMs may compromise the ability to discriminate debris-free from debris-covered ice slopes.

5) The semi-automatic method is better at discriminating clean ice from debris-covered ice compared than visual assessment, and at discriminating finer detail in patchy debris cover. Ideally, however, both methods should be applied in tandem and areas of disagreement targeted for more detailed investigation.

6) Where the semi-automatic method cannot be applied, the manual method should be performed independently by operators and areas of disagreement targeted for further study, to try to reduce the subjectivity in this method.

REFERENCES


Deline, P. & Orombelli, G. “Glacier fluctuations in the Western Alps during the Neoglacial, as indicated by the Miage moranic amphitheatre (Mont Blanc massif, Italy)” Boreas, vol 34, p.p. 456-467, November 2005


