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Ground temperature and permafrost distribution in Hurd Peninsula (Livingston Island, Maritime Antarctic): An assessment using freezing indexes and TTOP modelling

Alice Ferreira [a,*], Gonçalo Vieira [a], Miguel Ramos [b], Alexandre Nieuwendam [a]

[a] Centre for Geographical Studies, IGOT, Universidade de Lisboa, Portugal
[b] Universidad de Alcalá de Henares, Spain

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ABSTRACT

The Western Antarctic Peninsula region shows mean annual air temperatures ranging from −4 to −2 °C. Due to its proximity to the climatic threshold of permafrost, and evidence of recent changes in regional air temperatures, this is a crucial area to analyse climate-ground interactions. Freezing indexes and n-factors from contrasting topographic locations in Hurd Peninsula (Livingston Island) are analysed to assess the influence of snow cover on soil's thermal regime. The snow pack duration, thickness and physical properties are key in determining the thermal characteristics and spatial distribution of permafrost. The Temperature at the Top Of the Permafrost (TTOP) model uses freezing and thawing indexes, n-factors and thermal conductivity of the ground, as factors representing ground-atmosphere interactions and provides a framework to understand permafrost conditions and distribution. Eight sites were used to calculate TTOP and evaluate its accuracy. They encompass different geological, morphological and climatic conditions selected to identify site-specific ground thermal regime controls. Data was collected in the freezing seasons of 2007 and 2009 for air, surface and ground temperatures, as well as snow thickness. TTOP model results from sites located between 140 and 275 m a.s.l were very close to observational data, with differences varying from 0.05 to 0.4 °C, which are smaller than instrumental error. TTOP results for 36 m a.s.l confirm that permafrost is absent at low altitude and thermal offsets for rock areas show values between 0.01 and 0.48 °C indicating a small effect of latent heat, as well as of advection.

1. Introduction

Evidence of climate change in Antarctic Peninsula is widespread and its impacts on the cryosphere have been focus of a number of studies, but impacts on permafrost, active layer and the ground thermal regime are still scarcely understood (Bockheim et al., 2013, de Pablo et al., 2013; Guglielmin and Vieira, 2014; Oliva and Ruiz-Fernández, 2015; Vieira et al., 2010). Permafrost, as a subsurface phenomenon, is difficult to assess and monitor, and its distribution is controlled by complex microclimatic factors such as snow, surface radiation budget, soil physical proprieties, moisture and vegetation (Williams and Smith, 1989).

Research has shown that a strong warming in the western Antarctic Peninsula occurred since the 1950's with an increase of ci. 3.4 °C in the mean annual air temperature (MAAT) in Faraday/Vernadsky until the early 2000's (Turner et al., 2005, 2009; Vaughan et al., 2003). Turner et al. (2016) have shown that since the late 1990's the Antarctic Peninsula showed a small, but statistically significant cooling, especially during the austral summer. The region has been one of the world's climate warming hot spots, having experienced a 40% of decrease in sea-ice coverage in the Bellinghausen Sea (Ducklow, 2008; Stammerjohn et al., 2008) and disintegration of ice shelves (Cook and Vaughan, 2010), but clearly the recent findings show that climate trends may be more complicated than previous records have shown.

Permafrost warming and degradation have been widely characterized in many regions of the Northern Hemisphere following atmospheric warming (i.e. Christiansen et al., 2010; Isaksen et al., 2011; Romanovsky et al., 2010, 2012; Schuur et al., 2015; Smith et al., 2010, 2012). In Antarctica the impacts of climate change (e.g. temperature and snow cover) on permafrost and active layer thermal regimes are, however, still not well known and such is the case of the Antarctic Peninsula. This gap in knowledge is essentially due to a lack of systematic data collection, as pointed by Bockheim (2004) and Vieira et al. (2010). Bockheim et al. (2013) compared recent field observations with literature from the 1980's and report permafrost degradation at low altitude near Palmer station (64° 77′ S).

Research by Serrano and Lopez-Martinez (2000) and Vieira et al. (2007), among other authors, validates Bockheim's (1995) proposal

* Corresponding author at: IGOT - Instituto de Geografia e Ordenamento do Território/ Centro de Estudos Geográficos, Universidade de Lisboa, Rua Branca Edmée Marques, 1600–276 Lisboa, Portugal.
E-mail address: alice.ferreira@campus.ul.pt (A. Ferreira).

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that the northern limit of continuous permafrost in Antarctica occurs in the Antarctic Peninsula and contiguous islands (Bockheim et al., 2008), with permafrost being absent close to sea-level in the South Shetlands (Serrano and Lopez-Martinez, 2000). Vieira et al. (2010) note that the continuous permafrost boundary in the Antarctic Peninsula region is to be found at a higher MAAT than those found in the Arctic, where the isotherms of −8 and −1 °C correspond roughly to the southern boundaries of continuous and discontinuous permafrost (Brown and Péwé, 1973).

The thermal regime of the active layer and permafrost depends on the atmospheric forcing but is also controlled by ground surface conditions, mainly by snow, its thickness, rhythm and thermal characteristics, which control energy fluxes between the ground and the atmosphere (Goodrich, 1982, 1978; Ishikawa, 2003; Lachenbruch, 1959; Romanovsky and Osterkamp, 2000). Such energy fluxes are especially difficult to characterize in alpine and polar maritime regions with rough topography, which increase the influence of local factors and particularly of snow conditions and solar radiation (Harris et al., 2003; Karunaratne and Burn, 2003; Klene et al., 2001; Taylor, 1995). Snow shows insulating properties acting as a buffer to heat losses in winter due to low thermal conductivity, while during the summer, high albedo and emissivity, induce cooling at the ground surface (Ramos and Vieira, 2009; Zhang et al., 2005).

Monitoring permafrost evolution and spatial distribution is a key approach for studying climate change, since permafrost regions are very sensitive to changes in air temperature and precipitation. Warm permafrost in the discontinuous and sporadic permafrost zones are most sensitive to climate change (André and Anisimov, 2009). Since it is impossible to map permafrost distribution by direct observations in large areas, several modelling approaches are used to estimate it, as well as its thermal state. Ground freezing and thawing indexes have been widely used for modelling permafrost distribution (Nelson and Outcalt, 1987), active layer thickness (Nelson et al., 1997; Shiklomanov and Nelson 1999, 2002; Zhang et al., 2005) and also for engineering applications in cold regions (Frauenfeld et al., 2007). Such indexes integrate the temporal changes in air-ground surface temperature fluxes, allowing for seasonal analysis and comparisons between sites (Klene et al., 2001). One of the most used indexes is the n-factor, which is the ratio between soil and air freezing indexes calculated separately for the freezing and thawing seasons (Carlson, 1952; Lunardini, 1978), designed to evaluate the degree of atmosphere and soil coupling, accounting for heat flux. N-factors are frequently used as a representative value of the joint insulating effects of vegetation, organic matter in the soil surface and snow conditions (Karunaratne and Burn, 2004; Lunardini, 1978; Throop et al., 2012).

The technological advances of temperature sensors fostered the research on the variability of n-factors in natural systems, especially concerning studies of thermal regimes in the active layer and permafrost (Burn and Smith, 1988; Karunaratne and Burn, 2003, 2004; Klene et al., 2001; Taylor, 1995, 2000; Throop et al., 2012) and further developments of modelling studies in permafrost areas (Juliussen and Humlum, 2007). One of the most successful methodological approaches to model the spatial distribution of permafrost is the empirical-statistical method of the Temperature of the Top of Permafrost (TTOP), developed by Smith and Riseborough (1996, 1998, 2002), linking the TTOP to climate through seasonal surface heat transfer functions and subsurface thermal proprieties (Henry and Smith, 2001). For such purpose, a relationship was established between an analytical model of the thermal offset effect, issuing from variations in the ground thermal properties, and the n-factors (Riseborough, 2008; Romanovsky and Osterkamp, 1995; Smith and Riseborough, 1996, 2002).

The focus of this paper is contributing to identify the microclimatic and topographical controls on ground surface temperatures (GST), as well as the role of snow cover and its significance for permafrost distribution in Hurd Peninsula (Livingston Island) using TTOP modelling. Hurd Peninsula is one of the areas of the Maritime Antarctic with a denser network of permafrost and active layer boreholes. MAAT varies between −2 and −3 °C and the area is in the boundary between continuous and discontinuous permafrost, being key for studying climate change effects on permafrost.

2. Study area

The South Shetland Islands are a mountainous and extensively glaciated archipelago, where permafrost occurs in bedrock and ice-free areas near the coast (Bockheim and Hall, 2002). Livingston (62° 39′ S, 60° 21′ W) is the second largest island in the South Shetlands archipelago and shows ci. 10% of its surface ice free, mainly in the lower areas of numerous peninsulas, many of them showing rugged mountain terrain. This study focuses on the ice-free areas of the north-western part of Hurd Peninsula in the vicinity of the Spanish Antarctic Station Juan Carlos I (SAS) and Bulgarian Antarctic Station St Kliment Ohridsky (BAS).

Hurd Peninsula is ci. 20 km^2 and culminates at 406 m a.s.l. at Moores Ridge near its southern tip. Three geomorphological settings are of significance to present-day periglacial morphodynamics and permafrost distribution in Hurd Peninsula: i. Areas with till and well-developed moraine ridges in recently deglaciated valleys; ii. Areas with widespread angular debris mantling slopes and interfluves, generally showing submetric to metric thicknesses, where deglaciation occurred earlier in the Holocene, and iii. Widespread areas with bedrock outcrops, either corresponding to convex terrain features (summits, saddles and knobs) or to cliffs and steep slopes. The most widespread periglacial landforms are stone-banked lobes, while other periglacial features include patterned ground, especially in the higher, wind exposed interfluves (e.g. non-sorted circles at Reina Sofia Hill) (López-Martínez et al., 1992, Vieira and Ramos, 2003). Talus-derived rockglaciers (e.g. Johnsons Ridge) and moraine-derived rockglaciers (e.g. False Bay, Hurd rockglacier) are present almost down to sea level (Hauck et al., 2007; Reis et al., 2015; Serrano and Lopez-Martinez, 2000; Serrano et al., 2004). Erosion surfaces with probable marine origin occur between 60 and 200 m a.s.l. (John and Sugden, 1971). Below 20 m a.s.l. along the coast the prevailing landforms are raised beaches (Curl, 1980; Hall and Perry, 2004; López-Martínez et al., 1992).

The bedrock of the study area is dominated by the Miers Bluff Formation, a low-grade metamorphic turbidite sequence with alternating layers of quartzite and shales, with conglomerates and breccias in some areas (Arche et al., 1996; Hobbs, 1968; Smellie et al., 1984, 1995). The age of this formation has been subject of debate but is currently attributed to the Late Cretaceous (Pimpirev et al., 2006). In the area of the Bulgarian Antarctic Station, the Hesperides Point Pluton occurs, forming a small stock composed of gabbro, diorites and quartz-diorites (Kamenov, 1997, 2008). The area is also cross-cut by numerous dykes (Kamenov, 2008) originating in six events dating from the Paleocene to Late Eocene, most showing tholeiitic affinity (Kraus et al., 2008).

Contrary to the Antarctic continent, the islands in the west Antarctic Peninsula are affected by eastward cyclonic depression fluxes providing more precipitation and less severe temperatures (Bockheim and Hall, 2002). The climate at sea level is cold maritime with frequent summer rainfall and a moderate annual temperature range, reflecting the influence of the circum-Antarctic low-pressure system (Simonov, 1977; Styszynska, 2004). Annual precipitation is ci. 500 mm at sea level. Annual air temperatures between 2000 and 2006 at the Spanish Antarctic Station Juan Carlos I (15 m a.s.l.) varied from −3.2 °C to −1.5 °C with a MAAT of −1.9 °C (Ramos et al., 2008a). At Reina Sofia Peak (275 m a.s.l.), the MAAT for 2003–2006 was −4.2 °C (Ramos et al., 2008b). Long-term snow cover data is absent for the study area, but ground surface temperature data indicate a high interannual variability of snow, as shown by de Pablo et al. (2016) for Byers Peninsula on the SW tip of Livingston Island. Snow covers the ground from April to December, with decimetrical thickness (10–40 cm). Shallower depths (10–20 cm) are found in wind swept sites at high elevation, while sheltered and low elevation sites can accumulate over 80 cm of snow.

Previous research on permafrost distribution in Livingston Island has been based on geomorphological evidence (Serrano and Lopez-Martinez, 2000; Vieira and Ramos, 2003; López-Martínez et al., 2012), ground temperature monitoring in shallow boreholes (Ramos and Vieira, 2003; Ramos et al., 2008a,b; Vieira et al., 2010) and geophysical surveying (Hauck et al., 2007). Geomorphological and geophysical surveys indicate that permafrost occurs immediately above sea-level associated to ice-cored moraines and rock glaciers, but in bedrock its identification is difficult and probably only occurs at higher altitude. Borehole data and excavations at Reina Sofia Hill (275 m a.s.l.) show the presence of permafrost with over 25 m depth (Ramos et al., 2009). Mean annual ground temperatures measured since 2000 (at depths of 15, 25, 40, and 90 cm in a 1.1 m borehole) varied between −2.6 and −2.1 °C. The active layer thickness, based on direct observations in pits and temperature data, was between 70 and 90 cm (Ramos et al., 2008a).

3. Methods

3.1. Monitoring sites

Air and ground surface temperatures were examined for eight sites in Hurd Peninsula ranging from 15 to 275 m a.s.l (Figs. 2 and 3 and Table 1): five in the vicinity of the SAS (Reina Sofia Hill - RSH, CR - Collado Ramos, Skidoo Hut - SH, Incinerator - INC and New Incinerator - NI) and three near the BAS (Papagal - PAP, CALM and Meteorological Station - MET). The sites spread across an altitudinal gradient and occur in different morphological settings and wind exposures and therefore also different snow conditions. The sites are flat in order to minimize slope effects on incoming solar radiation and show no vegetation cover. This setup allows for assessing the effects of altitude and snow cover on ground temperatures. The ground surface is normally covered by frost shattered debris, mixed with sandy-silty material forming a diamicton, which is usually shallow (0.1-1 m). No characterization of the albedo of bare ground has been done, but the colour does not vary significantly and we assume that in such a high cloudiness environment, where ci. 90% of the days are cloudy, the variability in albedo does not significantly affect warming as in other drier regions, such as Continental Antarctica. Bedrock outcrops are frequent and some were used to install boreholes. The sites were installed in the framework of monitoring programs of Portugal and Spain and the setup follows slightly different protocols. In this paper, only the records from 2007 and 2009 are analysed, since several loggers malfunctioned in 2008. Air and ground surface temperatures (GST) were measured at five sites in 2007 (RSH, CR, SH, INC and NI) and at seven sites (RSH, CR, INC, NI, MET, CALM and PAP) in 2009 (Table 1). (See Fig. 1.)

3.2. Ground surface temperatures

Two types of single-channel miniloggers were used for GST: TinyTag 2 (Gemini UK) and Thermochron iButtons DS1922L, both showing an accuracy of 0.3 °C. At RSH and INC, the TinyTag 2 was at 5 cm depth inserted in a PVC tube in boreholes and temperature was recorded at 1-h intervals. At CR the setting was similar, but the loggers were installed at 2.5 cm depth. At NI, CR, SH, MET, CALM and PAP, GST were measured at 2 cm depth, using buried iButtons fixed to a 15 × 15 cm high diffusivity aluminium plate measuring at four hour intervals (Table 1). Despite the differences in setup, we consider that the data is comparable, allowing for a good approximation of ground freezing and thawing indexes. This is valid since Livingston Island is a high cloudiness environment, with high humidity and a low diurnal thermal range, which induces a small ground temperature gradient and slow temperature changes. The differences in measuring interval do not impact the data, since average daily temperatures are used for the degree-day calculations.

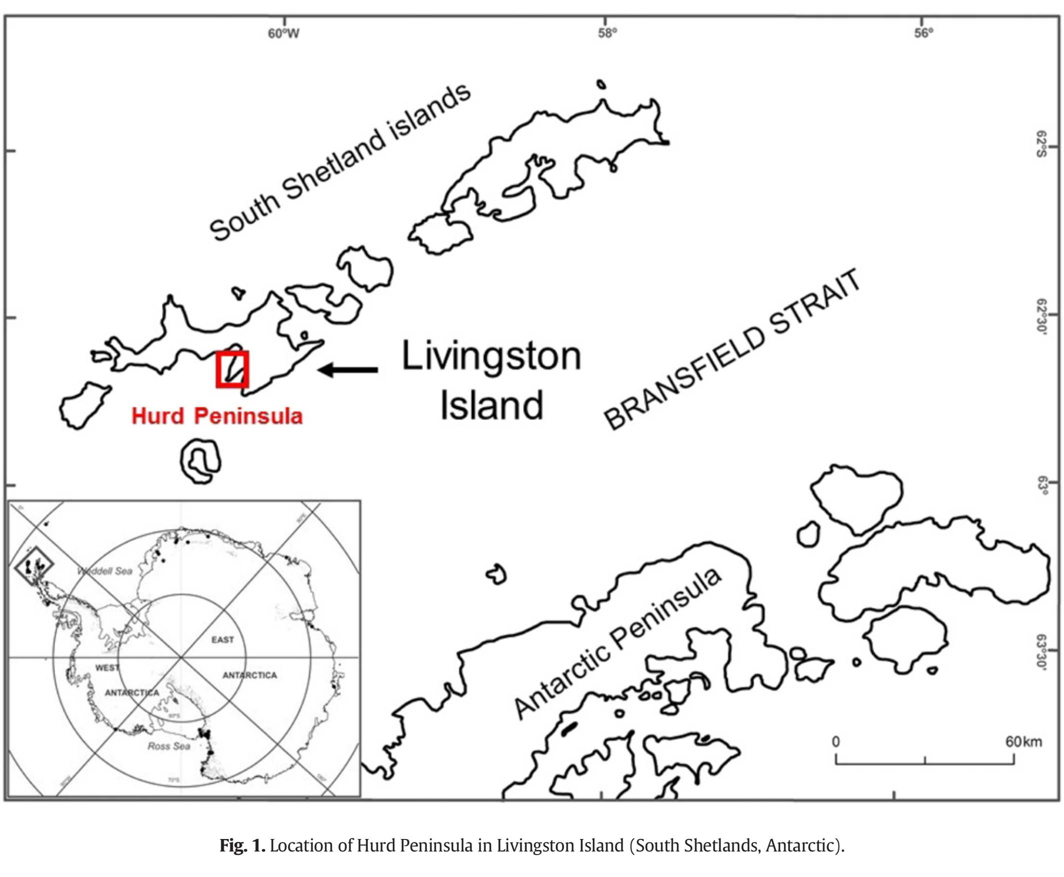

Fig. 1. Location of Hurd Peninsula in Livingston Island (South Shetlands, Antarctic).

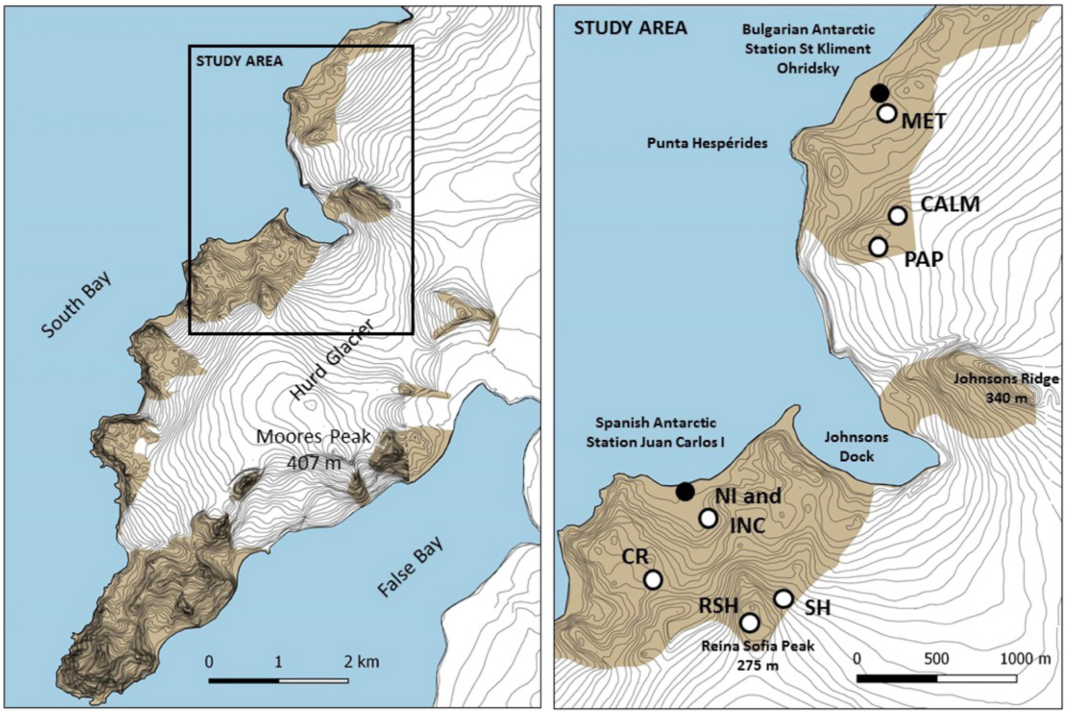

Fig. 2. Location of the study area and monitoring sites in Hurd Peninsula.

3.3. Mean annual ground temperatures

Mean Annual Ground Temperatures (MAGT) were calculated for RSH, MET, PAP and CALM, where active layer and permafrost boreholes exist. At RSH, borehole temperatures are measured with one-hour interval down to 25 m using negative temperature coefficient-resistors type YSI 44031, with an accuracy of ci 0.1 °C (Ramos et al., 2009). At MET borehole temperature recordings are made using Hobo Pendant mini-

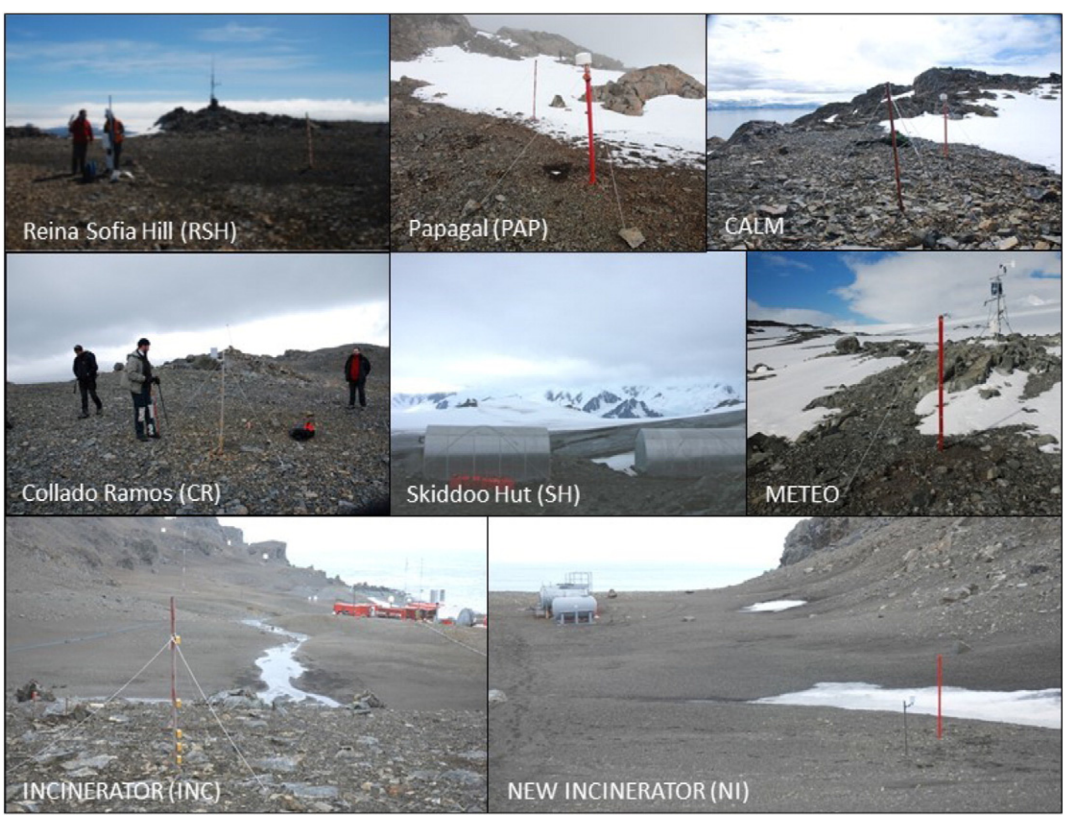

Fig. 3. Overview of the monitoring sites in Hurd Peninsula.

Table 1
Geographical setting and data availability for the monitoring sites in Hurd Peninsula.

Sites	Alt (m asl)	Geomor-phological setting	Wind exposure	Geology/Surface material	Environmental variables				Periods with data	
					AT	GST	MAGT	Snow (2009)	2007/08	2009/10
RSH	275	Slope top near summit	Very high	Shallow diami-cton	x	x	x	x	10/02–23/01	16/01–23/12
PAP	152	Wide ridge	High	Andesite	x	x	x	x	–	01/02–04/01
CALM	140	Wide flat interfluve	Mode-rate	Shallow diami-cton	x	x	x	x	–	
SH	140	Flat step in slope	Mode-rate	Shallow diami-cton	x	x			10/02–22/01	–
CR	117	Wide flat interfluve	High	Shallow diami-cton	x	x	x	x	10/02–22/01	14/01–22/12
MET	36	Small rock knob	High	Quartzite	x	x	x	x		01/02–13/01
INC	35	Rock step in slope	Low	Quartzite	x	x	x	x	10/02–22/01	14/01–22/12
NI	15	Slope foot	Very low	Silt/coar-se sand	x	x	x	x	10/02–22/01	14/01–17/12

loggers at one-hour interval down to 8 m. At PAP (4 m) and CALM (5 m) boreholes, ground temperatures are recorded at four hour intervals with Thermochron iButtons DS1922L. Air temperature was measured at 1.5 m from the ground with 1-h interval using TinyTag Plus 2 miniloggers.

3.4. Snow thickness

Snow thickness was measured at INC and NI in 2007 and at all sites in 2009. Following Lewkowicz and Bonnaventure (2008), Thermochron iButtons DS1921G mounted on wooden stakes at heights 2.5, 5, 10, 20, 40, 80 and 160 cm, were used to determine site-specific snow depth. The procedure allows to estimate the presence of snow, as well as its approximate thickness by comparing temperatures at different levels along the stake, with the air temperature above (Lewkowicz et al., 2012). The snow thickness values correspond to the height of the topmost temperature sensor buried by snow. A value of 40 cm indicates that snow cover is over 40 but <80 cm thick.

3.5. Freezing and thawing indexes

In order to calculate the freezing and thawing indexes, the freezing and thawing seasons were defined as the periods when the mean daily temperatures were consistently below and above 0 °C, respectively. At all study sites, the air and ground freezing seasons started almost simultaneously, since snow cover at the end of the thawing season was negligible. The definition of the beginning and end of the freezing and thawing seasons is determined using the GST following Karunaratne and Burn (2004). The sum of the daily mean temperatures above 0 °C during the thawing season and the sum of the modulus of the daily mean temperatures below 0 °C during the freezing season defines the thawing air and soil (DDTa,s) and freezing indexes (DDFa,s), respectively (Klene et al., 2001).

Seasonal thawing (n_t) and freezing (n_f) n-factors have been calculated from air and ground surface data following Lunardini (1978):

$$n_t = DDT_s / DDT_a$$
$$n_f = DDF_s / DDF_a$$

where: n_t – thawing n factor, DDTs thawing degree-days of ground surface (Celsius degree-days); DDTa thawing degree-days of air (Celsius degree-days); n_f – freezing n factor, DDFs freezing degree-days of ground surface (Celsius degree-days); DDFa freezing degree-days of air (Celsius degree-days).

3.6. The temperature at the top of permafrost model

The Temperature at the Top Of Permafrost (TTOP) model, provides a functional framework of the climate-permafrost system, relating the influence of climate, terrain and lithological factors on the thermal regime and distribution of permafrost (Henry and Smith, 2001). The following formula sets the model:

$$TTOP = \frac{\frac{kt}{kt}\,\mathrm{nt\ DDTa} - \mathrm{nf\ DDFa}}{\mathrm{p.}}$$

where: nt - thawing n-factor, nf - freezing n-factor, kt - thermal conductivity of ground in thawed state, kf - thermal conductivity of ground in frozen state, DDTa - thawing degree-days of air, DDFa - freezing degree-days of air, and P - number of days in one year.

This empirical-statistical model has been widely used to assess climate change impacts on permafrost (Smith and Riseborough, 1996), to determine the climatic thresholds of permafrost (Smith and Riseborough, 2002), to map permafrost spatial distribution at a continental scale (Henry and Smith, 2001), at a regional scale in Colorado (Janke et al., 2012), Norway (Gisnås et al., 2013; Juliussen and Humlum, 2007), Alaska (Kerkering, 2008), and the Mackenzie region (Stevens et al., 2010; Wright et al., 2003) and Yukon, (Bevington, 2015).

TTOP analysis was made for INC, MET, CR, CALM, PAP and RS for 2009, which is the only year with borehole data allowing for model validation. In the current approach we neglect the presence of soils and deposits and use bedrock thermo-physical properties for the TTOP modelling. Moisture variability and phase change issues were not accounted for, a fact minimised since the boreholes with observational data used for validation were drilled in bedrock. This allows for assessing the quality of the modelling, but the results have to be interpreted with caution, due to the assumptions that were followed. Rock thermal conductivity data and its variation with depth was obtained in laboratory by Amaral (2011) and Correia et al. (2012) for all sites except INC and CR, where tabulated values were used (Schön, 1996).

One strong point in the present TTOP modelling approach was that the variables in the model were effectively measured, while several other studies use a smaller number of measured input values (Gisnås et al., 2013; Janke et al., 2012; Juliussen and Humlum, 2007) or even tabulated values (Henry and Smith, 2001; Stevens et al., 2010; Wright et al., 2003).

4. Results

4.1. Snow cover and ground surface temperature during the freezing season

Mean daily air and surface temperatures and snow thickness are shown in Figs. 4 and 5, for 2007 and 2009 respectively and freezing season parameters are represented in Table 2.

In 2007, snow thickness below 35 m altitude varied between 10 and 20 cm from the end of April to mid-July showing a small influence on the variation of GST. From there on, snow cover showed a thickness of 20 to 40 cm and GST variability was lower. At INC, GST ranged between −4 and −2 °C from mid-July to mid-September and from −2 to 0 °C throughout the rest of the freezing season. At NI the range was lower

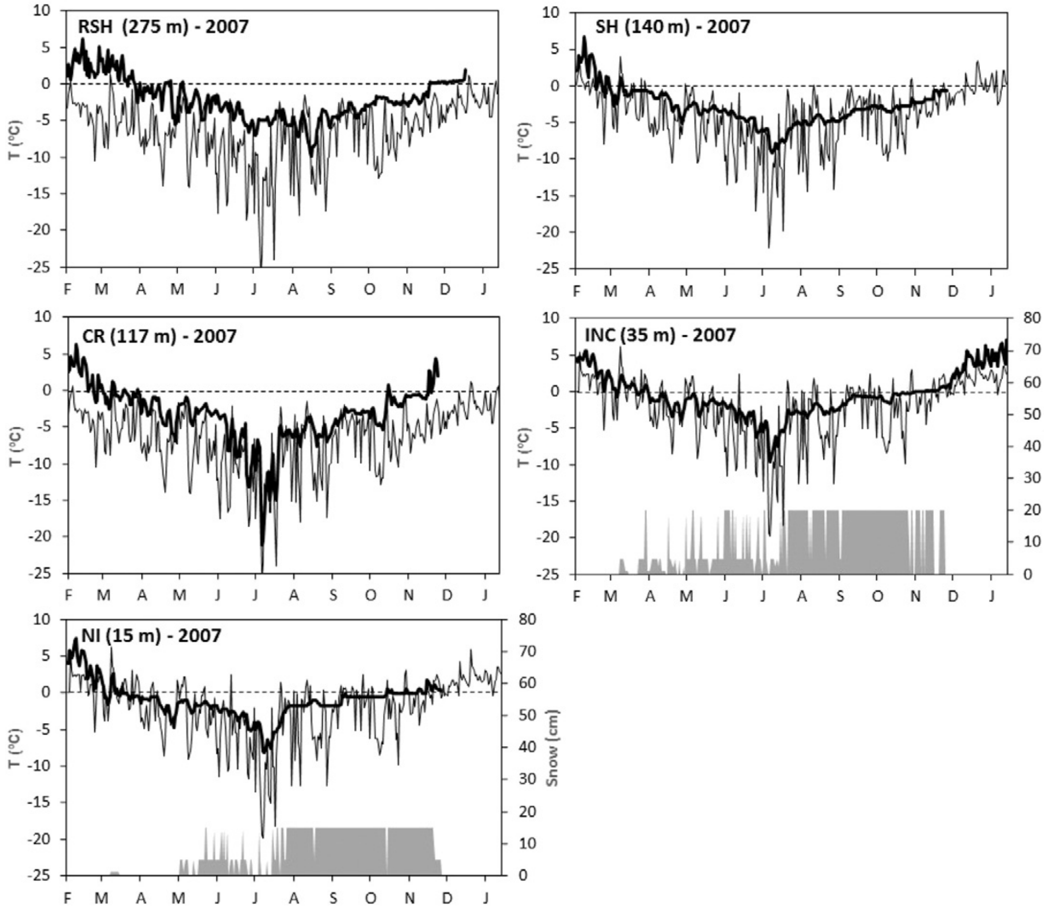

Fig. 4. Daily mean air and soil surface temperatures and snow thickness (ST) for monitored sites in Hurd Peninsula for 2007. Black solid thick line - soil surface temperatures, Black solid thin line – air temperature, grey solid line – snow thickness (cm).

and temperatures ci. 2 °C higher, despite the sites showing an altitude difference of only 20 m.

The mid-altitude sites show significant differences between them. At CR the coupling with atmospheric temperatures is high throughout almost all season except the last three months. At the SH, GST prevails over 3 °C warmer than air during most of the freezing season.

Snow thickness for 2007 at higher altitudes sites is not available, but ground temperature data suggests a smaller snow insulation and a stronger coupling with air. At RSH GST show a slow response to air temperatures since June, ranging from −2 to −7 °C, which could be the result of a thin snow cover.

Snow cover in 2009 showed smaller influence on the GST throughout the freezing season. In April the highest elevation sites (above 140 m) show a larger offset but only in June the differences between air and GST begin to accentuate, with differences ranging from 5 to 9 °C. At CR, CALM and INC in August, GST show a rhytm variability close to that of the air temperatures, probably due to a thin snow cover. In October the snowpack is thicker for all sites and the insulating effect is evident, especially at NI.

4.2. Freezing index variability

4.2.1. Air freezing

Air freezing indexes in 2007 were lower than in 2009. In 2007 the highest site (RSH) showed almost 2000 DDF and intermediate sites such as CR and SH showed 1657 and 1400 DDF, respectively. The lowest sites (NI and INC) showed ci. 900 DDF. In 2009 the air freezing index in RSH was over 500 DDF lower, a difference which is close to what occurred in the mid-elevation sites. The very low DDFa at NI was due to the burial of the air temperature logger under a very thick snow mantle.

4.2.2. Soil freezing

The freezing season began earlier in 2007 than 2009 as also did the onset of the snow pack (Fig. 5), giving origin to lower soil freezing indexes especially in CR (1074 DDF in 2007 and 772 DDF in 2009) and RSH (1042 DDF in 2007 and 811 DDF in 2009). As shown in Fig. 4, in 2007 the similarities between sites are provided by altitude.

Table 2 shows that the beginning of the freezing season tends to be later at the lowest altitude sites where it also shows a shorter duration. Exception to this is found when the soil is under a thick snow pack (e.g. SH in 2007 and NI in 2009). Above freezing days are also more frequent at the low altitude sites either in the air and soil, particularly at the ground surface.

The soil degree-days of freezing are systematically lower than air, with average values in the order of 484 DDF in 2007 and 272 DDF in 2009. With exceptional differences at the 275 m elevation in RSH and low elevation at NI the air index nearly (2007) and more than (2009), respectively, doubles, the freezing soil index values.

4.2.3. Altitude control on freezing indexes

Air freezing degree days show a strong variation with altitude, with differences of over 1000 DDF in 2007 and 750 DDF in 2009, in a range of only 260 m (Fig. 6). A strong linear correlation is found between DDFa and altitude for 2007 and 2009, with r^2 values of 0.88 and 0.92

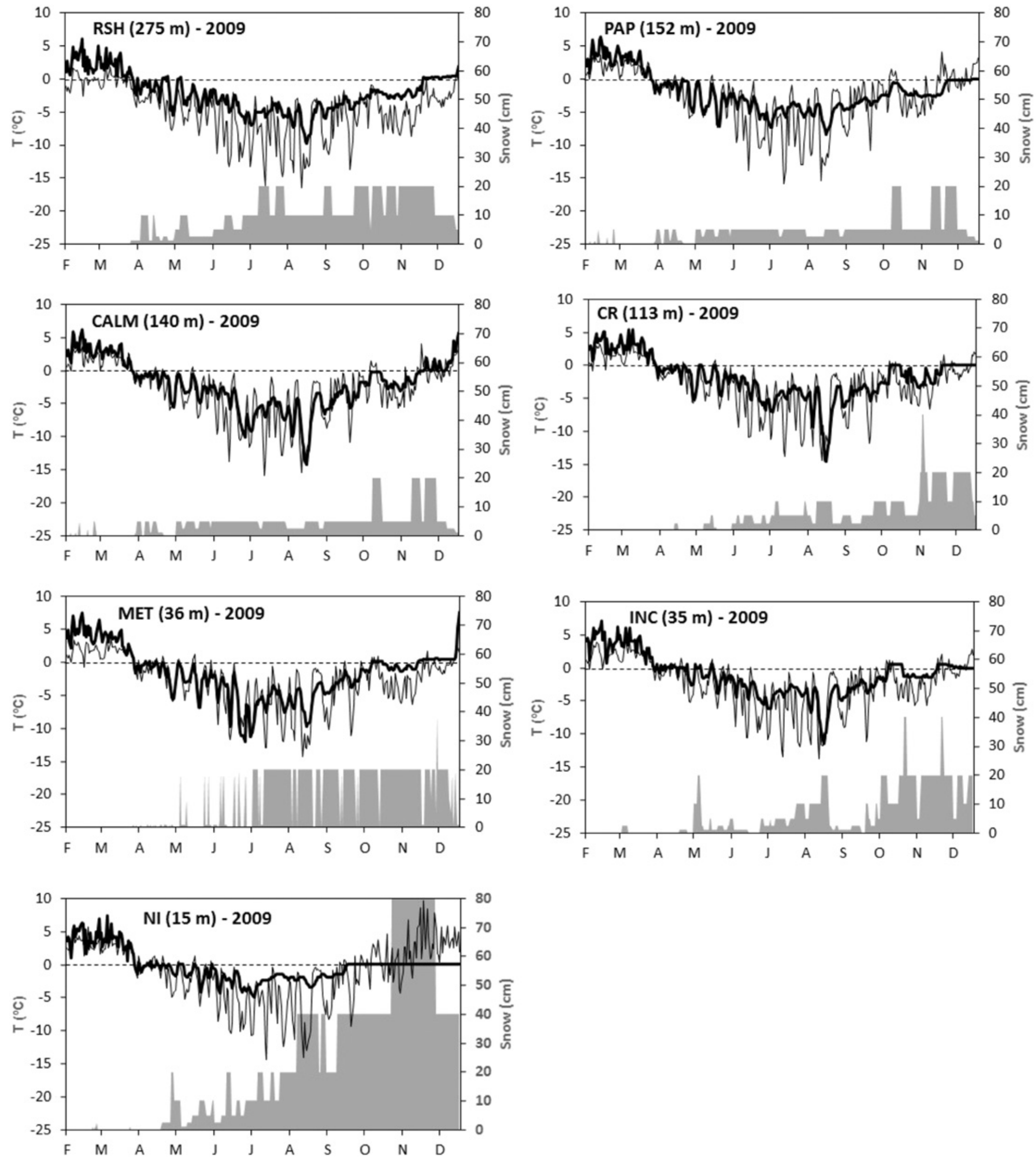

Fig. 5. Daily mean air and soil surface temperatures and snow thickness (ST) for monitored sites in Hurd Peninsula for 2007. Black solid thick line - soil surface temperatures, black solid thin line – air temperature, grey solid line – snow thickness (cm).

respectively, statistically significant with $p < 0.02$ for 2007 and $p < 0.0006$ for 2009 (Fig. 5). The correlation between DDFs and altitude shows lower statistical significance, with r^2 values of 0.67 for 2007 and 0.40 for 2009, with $p < 0.09$ and $p < 0.13$, respectively. The low significance is due to the existence of controls, other than altitude, for explaining soil temperature, such as snow cover. The altitudinal change

Table 2
Characteristics of the freezing seasons of 2007 and 2009 in Hurd Peninsula.

			Freezing season		Nr days >0 °C		DDT		DDF			
	Sites	Alt. (m asl)	Period	Nr days	Air	Soil	Air	Soil	Air	Soil	nt	nf
2007	RSH	275	19/03–28/11	255	2	21	4	49	1969	1042	10.95	0.53
	SH	140	09/03–05/12	272	16	25	36	62	1400	915	1.72	0.65
	CR	117	10/03–25/12	261	10	35	11	85	1657	1074	7.69	0.65
	INC	35	09/04–12/11	212	91	79	172	313	890	487	1.82	0.55
	NI	15	27/03–16/11	235	86	55	159	127	901	434	0.80	0.48
2009	RSH	275	28/03–19/11	236	40	113	21	145	1434	811	6.79	0.57
	PAP	152	26/03–17/11	235	103	69	137	214	1123	783	1.55	0.70
	CALM	140	27/03–19/11	238	101	93	157	241	1123*	880	1.54	0.78
	CR	117	28/03–19/11	237	90	119	133	227	1017	772	1.71	0.76
	MET	36	28/03–13/11	231	96	129	140	318	965	771	2.27	0.80
	INC	35	01/04–17/11	230	110	143	189	312	890	583	1.65	0.65
	NI	15	28/03–17/12	265	166	185	446	339	686	294	0.76	0.43

* Data from PAP (located in the vicinity at similar altitude).

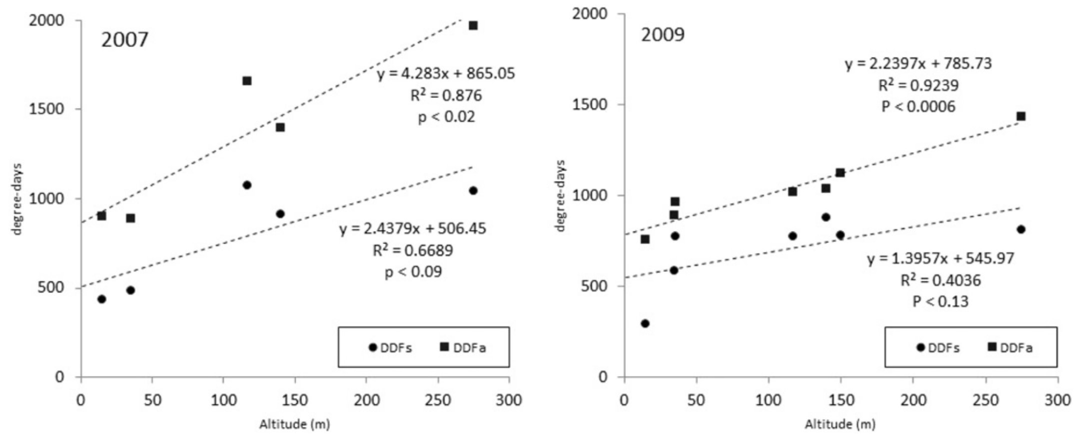

Fig. 6. Regression analysis between DDFa, DDFs and altitude for 2007 and 2009.

rate for DDFa was over 400 DDF/100 m for 2007 and ci. 300 DDF/100 m for 2009. DDFs showed a rate of over 200 DDF/100 m for both years.

4.2.4. N-factors

The highest n-factor was recorded at MET with 0.8 in 2009, a value close to those shown at mid elevation sites, but much larger than the other low elevation sites in the same year (INC – 0.65 and NI – 0.43). In 2007 the highest values were recorded at mid-elevation sites (SH and CR with 0.65). In 2007 RSH and NI showed n-factors of 0.53 and 0.48, and in 2009, 0.57 and 0.43, respectively. These sites have consistently showed the lowest n-factors. In 2009 the highest n-factors occurred at the mid elevation sites (CR, CALM and PAP) varying from 0.70 to 0.78.

4.3. TTOP modelling and validation

TTOP were calculated for six locations and only for 2009, which is the period with snow cover data (Table 3). The results are estimated values of mean annual temperature at the permafrost table, with the lowest values indicating higher probability of existence of permafrost. A general decrease in modelled TTOP with altitude is observable, with values ranging from −0.57 (TTOP min at INC) to −1.83 °C at (TTOP max at RSH).

Values of modelled TTOP at RS are the lowest and differ from observations between 0.42 and 0.84 °C (Table 3). The thermal offsets were not directly calculated for PAP and CALM since the boreholes do not reach the permafrost table. However, the maximum annual ground temperatures at the deepest sensor in both boreholes show values close to 0 °C, suggesting the possible presence of permafrost slightly below (Fig. 7). For this two cases, TTOP was estimated by extrapolation and the results are very close to the modelled TTOP, with differences of 0.03 at CALM and 0.1 °C at PAP (Table 3).

5. Discussion

5.1. Effects of snow cover and altitude on the ground surface temperatures

The temporal and spatial variability of n-factors is caused largely by the changeability of surface conditions in natural environments (Klene et al., 2001). Freezing n-factors decline with increasing snow depth as the atmosphere cooling is restricted by the insulating effects of the snow pack (Karunaratne, 2002; Klene et al., 2001; Riseborough and Smith, 1998).

The analysis of the freezing periods for 2007 and 2009 show a strong control of elevation in snow duration and thickness, with a significant variability during the freezing season for all sites above 35 m (Figs. 3 and 4). At the lower altitudes, snow was more stable and lasted longer due to wind shelter and shadow effects. 2007 showed lower DDFa, DDFs and also lower n-factors for most sites, illustrating a strong buffer effect of the snow cover. Lower altitude sites (INC and NI) showed smaller DDFs, with snow cover showing an enhanced and longer insulating effect. The effect of wind on snow cover is emphasised at MET, where the n-factor showed the lowest value of all. This was due to the rock knob position of the site, with strong exposure to wind and scarce snow cover. Intermediate elevation sites (CR and SH) showed higher coupling between air and soil, especially during the middle of the freezing season, particularly at CR, and indicate a lower insulating effect of snow cover, a fact also related to snow redistribution by wind, associated with the local geomorphic setting (flat or convex) that facilitates snow removal. The highest site, Reina Sofia showed an intermediate n-factor, controlled by snow accumulation since the site is located close to the summit, but on a slope site with more snow than the intermediate sites, such as CR.

The results confirm that snow cover is a major factor controlling the air-ground interface, since vegetation is absent. Timing and duration of snow have a direct impact on surface albedo, which in turn affects the surface energy balance, thus influencing the thermal regime of active

Table 3
TTOP input parameters, modelling results and observational data for validation for the monitoring sites in Hurd Peninsula in 2009. TTOP, MAGST and Thermal offset in °C.

Site	K ratio	DDTa	DDTs	TTOP (min)	TTOP (max)	MAGST	Thermal offset (max)	Thermal offset (min)	Measured TTOP
RSH	0.92	21	145	−1.40	−1.86	−1.83	0.43	0.01	−2.24
PAP	0.92	137	214	−1.23	−1.55	−1.71	0.48*	0.16*	−1.3*
CALM	0.94	157	241	−1.52	−1.78	−1.88	0.36*	0.10*	−1.52*
CR	0.87	157	227	−1.16	−1.39	−1.58	0.42	0.20	–
MET	0.96	140	318	−1.02	−1.28	−1.30	0.28	0.02	–
INC	0.87	189	312	−0.57	−0.69	−0.78	0.21	0.08	–

* Values estimated by extrapolation.

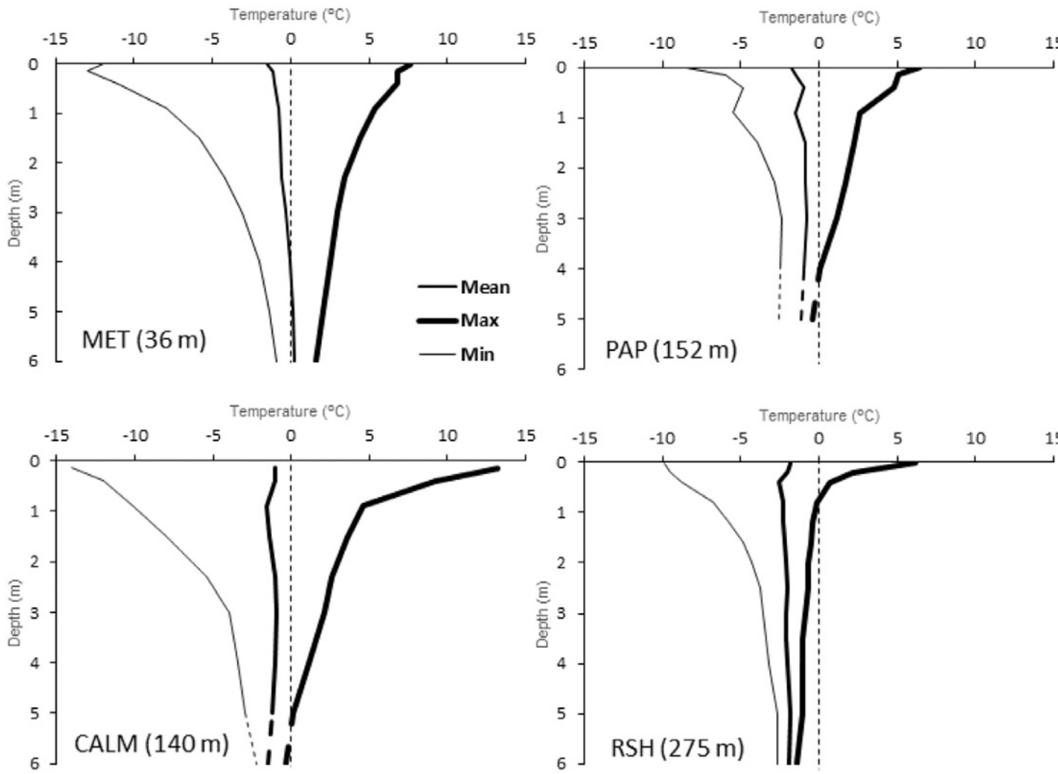

Fig. 7. Mean, maximum and minimum annual ground temperatures in 2009 at four boreholes in Hurd Peninsula and extrapolated mean values of temperatures at the permafrost table for PAP and CALM sites. All boreholes are installed in bedrock.

layer and permafrost. The major role of snow cover on the ground thermal regime has been shown for several sites in the Western Antarctic Peninsula, such as Adelaide Island (Guglielmin et al., 2014), Amsler Island (Wilhelm and Bockheim, 2016), Byers Peninsula (de Pablo et al., 2013, 2014, 2016; Oliva et al., 2016), and Deception Island (Goyanes et al., 2014). Hrbáček et al. (2016) have shown that in the Eastern Antarctic Peninsula snow influence on ground temperature is much smaller, especially due to the colder and drier climate conditions of the region.

5.2. TTOP modelling

5.2.1. Significance of the modelled TTOP results

The evaluation of the application of TTOP modelling to Hurd Peninsula, despite the limitations of using data from a single year, showed important results. The modelled TTOP for PAP and CALM showed differences of only 0.03 and 0.1 °C when compared to the values extrapolated from observations. Larger differences at RSH of 0.42 to 0.84 °C are probably related to the simplification of the modelling, since the presence of a surficial diamicton in the upper ci 1 m of the borehole was not included in the model. In Norway, Juliussen and Humlum (2007) found differences between modelled and measured TTOP below 1 °C and Bevington (2015) in Canada reported differences between 1 and 2 °C. This indicates that our TTOP results are promising for future applications arising from more numerous GST measurements, which may allow for a good approach for the spatialization of the TTOP.

5.2.2. Thermal offset

The thermal offset is defined as the difference between TTOP and the MAGST (Smith and Riseborough, 2002) and is mainly conditioned by changes in thermal conductivity (Burn and Smith, 1988). If moisture is present in the ground, the thermal conductivity is much higher when the site is frozen than when thawed, being a result of ice being four times more conductive than water (Smith and Riseborough, 2002). This fact allows heat to flow up more easily through the active layer in the winter than down in summer (Judge, 1973; Taylor, 2000; Throop, 2010). Given that the thermal conductivities that were used are for bedrock, where water content and latent heat fluxes are negligible, the resulting thermal offsets are low, with values from 0.01 to 0.48 °C and are close to offsets of 0.3 to 1.0 °C presented by Guglielmin and Cannone (2012) for Adelaide Island. Positive thermal offsets indicate MAGST < TTOP and negative thermal offsets indicate MAGST > TTOP (Bevington, 2015). Such differences may be the result of latent heat requirements for melting ice in the active layer (Hasler et al., 2014), or due to seasonal differences in thermal conductivity values (Bevington, 2015).

5.2.3. Contribution to the characterization of the spatial distribution of permafrost

Ground Penetrating Radar (GPR) measurements made in the vicinity of the CALM site at about 140 m a.s.l. by Schwamborn et al. (2008) showed a thaw layer of 40 cm in a volcanic sand and sandstones deposit. More recently, in the summers of 2009, 2012 and 2013 several electrical resistivity tomography (ERT) surveys were done close to CALM in an area with bedrock (Miers Bluff Formation) covered by frost-shattered debris (Correia et al., 2013). The results showed patches of either seasonally frozen ground or sporadic permafrost, which persisted throughout the years. The boreholes in bedrock at PAP and CALM with 4 and 5 m depth do not reach the permafrost table, although the extrapolation of temperature data suggests that permafrost may be present about 1 m below the borehole. The shallow thaw depth found by Schwamborn et al. (2008) and by Correia et al. (2013) may be linked to the influence of late lying snow in the ground, together with the lower thermal conductivity of saturated shallow surficial deposits. Furthermore, permafrost is known to occur at Reina Sofia Hill, at 275 m a.s.l, with temperatures at the zero annual amplitude depth of −1.8 °C (Ramos et al., 2009, Vieira et al., 2010). These observations, together with the TTOP model results suggest that 150 m asl is an altitudinal zone close to the continuous permafrost favour the presence of permafrost above about 150 m a.s.l.

6. Conclusions

One of the main objectives of this research was to determine GST variability controls throughout the freezing season. The main controlling factor was the combined effect of topography and wind influence on snow conditions, specifically thickness, timing and duration of the snow pack. The spatial variability of these factors conditioned GSTs during the freezing seasons of 2007 and 2009. Altitude also played a role, although less significant.

The TTOP model for RSH, as well as for PAP and CALM, where the values at the permafrost table were extrapolated, showed very good results. However, our study is constrained by two limitations: the number of boreholes used for validation is scarce, and only one year was studied, inducing limitations associated with the variability of interannual surface conditions derived from the input n-factor parameterization. Calculated thermal offsets were low, reflecting the thermal conductivity of bedrock.

Permafrost presence in bedrock at the lowest sites is very unlikely, since temperatures at 8 m depth did not reach the permafrost table and the modelled values for TTOP were high. This conclusion is consistent with observations from geophysical surveys. TTOP model values suggest a high probability of continuous permafrost in bedrock at 140 and 150 m elevation, although with an active layer of over 5 m. At mid altitude sites, such as CR (117 m a.s.l), TTOP values were close to those of the MET site and geophysical surveys showed that frozen ground was present only in particular conditions influenced by late lying snow (sporadic or patchy permafrost) (Hauck et al., 2007). Therefore, the altitudinal limit for continuous permafrost in bedrock is likely to be above 140–150 m, although with very thick active layers (>5 m).

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