Temporal and Spatial Variability of Winter Thermal Background Scenes

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ABSTRACT

This paper contrasts three winter backgrounds at a northern New England site in terms of surface temperature range, rate of temperature change, and spatial homogeneity. Ground cover surface temperatures usually are expressed as averages over several hours or days, which makes seasonal differences in thermal radiance apparent but obscures shorter-term variations that affect energy exchanges and infrared sensor systems. For this study, surface temperatures of the three winter ground covers are determined at half-hour intervals. The early winter background is a uniform grass cover following the end of the growing season. By late winter this has become a heterogeneous ground cover of thatch, new-growth grass, and exposed soil, which is a dynamic thermal background with strong daytime/nighttime differences. The midwinter snow cover is a stable thermal background because of its typically low rates of temperature change and low thermal radiance. It is shown that these three backgrounds produce distinctly different responses by passive infrared thermal devices (PIRs) that are sensitive to the spatial variability of changes (both magnitude and rate of change) in thermal radiance from the area being viewed. It is proposed that a PIR could be used to determine remotely the nature of the ground cover (grass, snow, grass–thatch–soil) and particularly to detect early and late winter transient snow covers.

Key words: Grass, ground cover, snow, soil, surface temperature, thermal radiance, thermal scene.

INTRODUCTION

The seasonal sequence of ground covers and site weather produces distinctly different winter thermal background scenes at the CRREL research site known as SOROIDS in South Royalton, Vermont. During the early winter, the dormant grass cover is a spatially homogeneous background that becomes a uniform sublayer to the eventual snow cover. The snow cover is succeeded by thatch intermixed with new-growth grass and exposed soil where the grass has been churned into mud during thaw periods. This is a thermally inhomogeneous background with no regularity to the spatial distribution of the thatch, exposed soil, or grass, all of which respond differently to heating and cooling in accordance with changes in insolation. Absorbed solar energy first dries the grass, thatch, and soil. Strong surface heating of the thatch (e.g., Bristow et al. 1986) and soil then follows, but the grass preferentially consumes the solar energy in evapotranspiration rather than through sensible heat exchange (e.g., Rosenberg 1974).

The snow cover is unique from the other backgrounds in that there is a natural limitation on its temperature, which cannot exceed 0°C. The thermal radiance of the snow essentially depends on the magnitude of the available solar energy only when there is no melting; increases in solar input after the onset of melting do not produce increases in the temperature of the snow surface. A corollary is that the maximum snow temperature need not necessarily occur during daylight hours, as, for example, when a warm air mass moves into the area at night. This can result in temporal variations in thermal radiance from a snow cover that are significantly different from the diurnal patterns characteristic of the other ground covers.

During the course of a SOROIDS winter, two thermal infrared sensor systems viewing a portion of the ground receive thermal radiance variously from dormant grass, snow, and the grass–thatch–soil ground cover. When the thermal systems are used to detect the movement of a person or vehicle through the area being viewed, a nonuniform change in thermal radiance from the ground cover can cause a false detection. This would be detrimental in applications such as battlefield surveillance or intrusion detection. When there is no intruder, each thermal system's response is a measure

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of the spatial variability of simultaneous changes in surface temperature throughout the area being viewed. If there is a predictable system response to temporal and spatial variations in thermal radiance characteristic of the typical ground covers at a site, then monitoring the thermal system becomes a means of remotely sensing changes in ground cover.

This paper shows that the seasonal sequence of ground cover and weather conditions at SOROIDS creates different thermal background scenes, in terms of surface temperatures and rates of change of temperature, which cause distinctly different responses by a thermal sensor system. It is proposed that such a thermal system could be used to remotely determine the nature of the ground cover (grass, snow, grass–thatch–soil), and particularly to detect early and late winter transient snow covers.

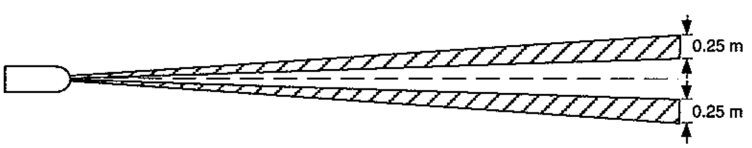
INSTRUMENTATION

The passive infrared thermal sensor systems, referred to as PIRs, are separated by 50 m on a roughly east–west line of sight, leading to a designation of PIR-East and PIR-West based on location (Fig 1). Each sensor system has two infrared detectors (8–14 μm spectral band) with different but closely spaced fields of view (Lacombe and Peck 1992). A thermal detector's field of view is a triangular sector of ground 50 m long and ~0.25 m wide at its farthest extent, or ~6.2 m^2 in area. The output of the two detectors is wired in a parallel–opposed manner so that a simultaneous, similar change in received thermal radiance at the two detectors produces no net response by the PIR, while a differential change in input to the two detectors causes a response determined by the magnitude and rate of change of the received thermal radiance. The net result of the PIR's detection and processing of a change in thermal radiance is expressed as a voltage that is a measure of the dynamics of the thermal scene. Each PIR's voltage is monitored with a data logger. The voltages are sampled at 8-Hz frequency, and maximum, minimum, and average values for the preceding 30 minutes are recorded every half hour. The voltage change is positive or negative, depending on whether the radiance change is high to low or low to high.

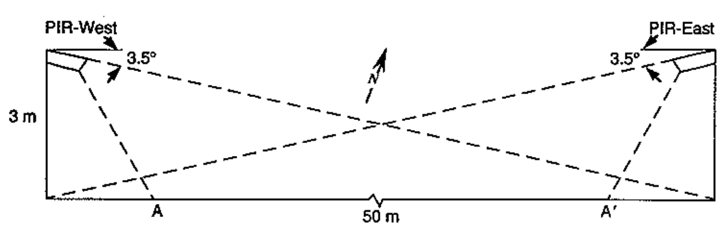
The apparent surface temperature of each background is calculated from the upwelling longwave radiation (3–50 μm) as measured with a downward-looking Eppley pyrgeometer. The radiation is averaged over 30 minutes and recorded every half hour. The field of view of the pyrgeometer is nominally a 180° hemisphere, but radiance received from an approximately 90° arc about the hemisphere's axis dominates. For a 1.5-m instrument height, this corresponds to a 1.5-m-radius circular area on the ground surface. The equation used to convert radiation (W/m^2) to temperature (°C) is a linear regression fitted to Planck's equation with the assumptions of 'black body' emission by the ground cover and flat, 100% response over the 3–50 μm spectrum (J. Lacombe, pers. com.). The actual surface temperature could be slightly different depending on the emissivity of each portion of the surface; however, emissivities of soils (0.90 to 0.98, depending on color and moisture content) and grass (0.90 to 0.95, depending on length [Oke 1987]), as well as snow (~0.99 [Warren 1982]), are high. The emissivity of the thatch is not known. The maximum error in calculated surface temperatures is estimated to be ~3°C for winter conditions and ~4°C for summer conditions, based upon the pyrgeometer response and the derivation of the equation used.

Incident solar radiation (30 min average) in the 0.3–3 μm bandwidth is measured with an Eppley pyranometer at a height of 1.5 m and includes the direct component of sunlight and the diffuse component of skylight.

Snow depth is determined automatically every 10 minutes with an acoustic snow depth sensor; the maximum and average depths are recorded every half hour. Air temperature and wind speed (both average and gust) at a height of 2 m are recorded every 30 minutes. Soil temperatures are determined with buried copper–constantan thermocouples.

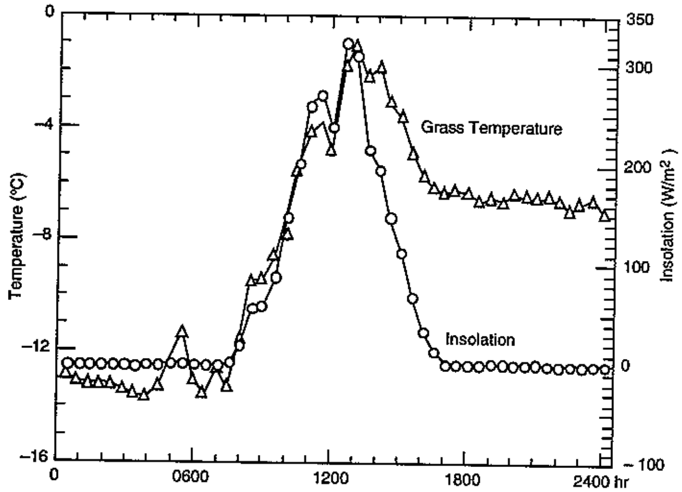


a. Top view showing the individual fields of view of a single PIR's two thermal detectors.

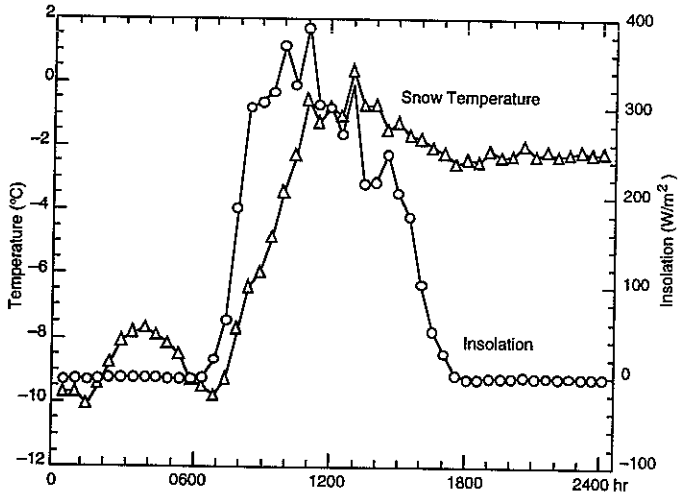


b. Side view showing the coverage of each PIR. The portion of the ground surface between A and A' is common to both PIRs.

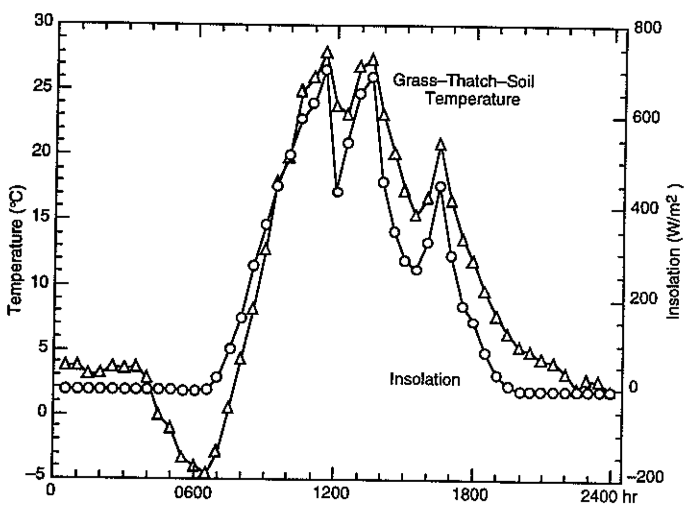
Figure 1. East and West PIRs at the South Royalton site.



a. Dormant grass cover on 10 January 1993.



b. Snow surface on 24 February 1992.



c. Grass–thatch–soil surface on 10 April 1992.

Figure 2. Time series records of ground surface (calculated) temperature and incident solar radiation (30-min average).

DIURNAL TEMPERATURE VARIATIONS OF THE BACKGROUND SCENE

Examples of the apparent surface temperature of the three winter ground covers are given in Figure 2, together with the incident solar radiation. The relative warming/cooling of a ground cover is dependent on the variation in magnitude of the insolation (with the exception of a snow cover, as noted above). The actual temperature distribution is determined by the heterogeneity of the ground cover and by the associated surface albedos for solar radiation (Table 1) and the range in thermal diffusivities of the ground-cover components. The lower its solar albedo, the more solar radiant energy a material absorbs. If the energy is not consumed in a phase change (latent heat of melting or evaporation) or readily dissipated into the underlying soil, then the ground cover heats and its surface temperature increases.

On 10 January 1993 (Fig. 2a), the ground cover was dormant grass with no snow cover present. Morning increases in the surface temperature of the grass cover were closely timed to increases in the incident solar radiation. Radiative cooling of the grass cover in the afternoon reduced its surface temperature to ~−6.5°C, 7°C warmer than before the morning onset of radiant-energy heating. Grass and air temperatures remained stabilized at −7°C overnight.

On 24 February 1992 (Fig. 2b), a snow cover 2–4 cm deep covered the site. Insolation (maximum half-hourly average 390 W/m^2) was typical for a SOROIDS winter day in the absence of snowfall or rainfall. Increases in the apparent surface temperature of the snow lagged the onset of insolation by 1.5 hours. On this sunny, calm (half-hourly wind gusts <3.5 m/s) day, the air temperature rose to above 0°C and the snow surface approached melting. The air and snow surface remained warm through the evening because of regional weather patterns.

On 10 April 1992 (Fig. 2c), the ground cover was new-growth grass poking through thatch and soil. Strong fluctuations in insolation because of passing clouds are evident in the daytime variation of surface temperature. The grass–thatch–soil surface was warmer than the underlying soil surface throughout the daylight hours, and it cooled below 0°C at night.

The diurnal range in apparent surface temperature of the grass–thatch–soil ground cover was 33°C on 10 April, compared with 13°C for the dormant grass on 10 January and 11°C for the snow cover on

Table 1. Surface albedos for solar radiation (after Lunardini 1981).

Surface	*Munn (1966)*	*Porkhaev (1959)*	*Van Wijk (1966)*	*Budyko (1974)*
Fresh snow	0.7–0.95	0.85	0.80–0.85	0.8–0.95
Old snow	0.7			
Melting snow			0.30–0.65	
Wet grass in sun	0.33–0.37	0.28		
Wet grass, no sun	0.14–0.26			
Dry grass	0.15–0.25	0.19	0.16–0.19	
Green grass			0.16–0.27	
Light-colored bare soil		0.35		
Dark-colored bare soil		0.15		

24 February. Solar warming during daylight hours and strong radiative cooling at night together can cause a variation in apparent surface temperature on the order of 29 to 45°C for this ground cover under April and early May weather conditions. This diurnal range in surface temperature is greater than or comparable to that of a lush grass cover in summer—23–37°C (Peck 1993)—when higher and protracted insolation can cause the grass surface to be warmer during daylight than the grass–thatch–soil of April, but the nighttime temperatures are also higher and so the diurnal range is less. In contrast, during snow-free periods in December and January, the dormant grass cover experienced a 9–13°C diurnal range in surface temperature.

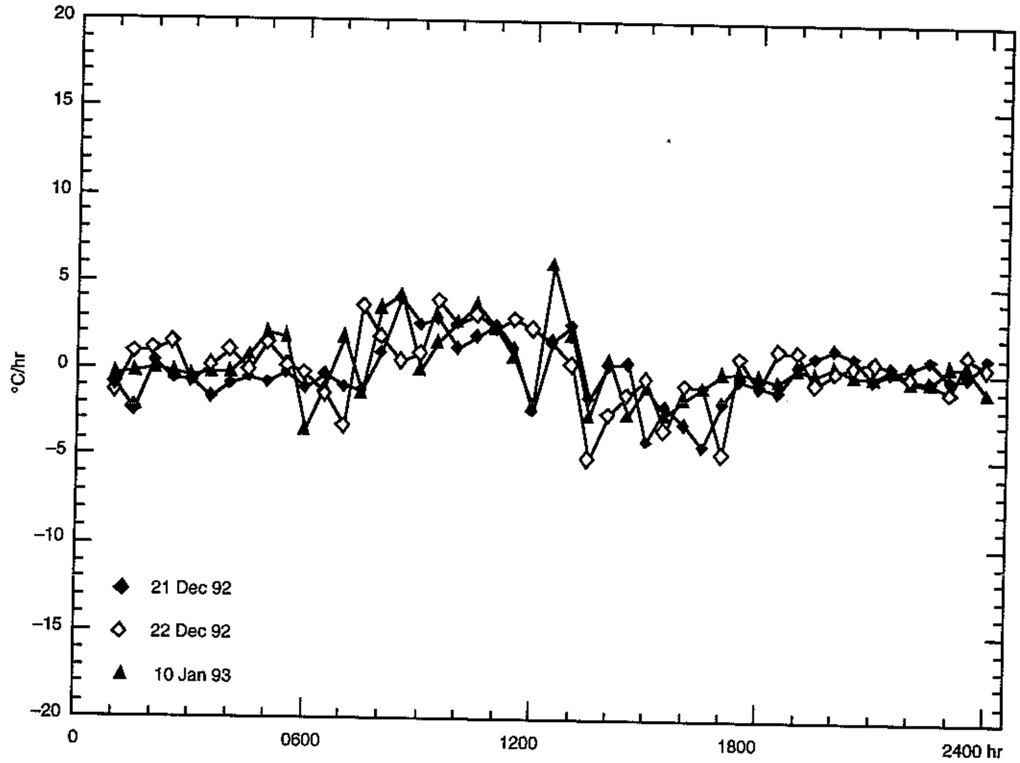
DIFFERENCES IN RATES OF CHANGE OF SURFACE TEMPERATURES

Apparent surface temperatures on 20 days between February 1992 and March 1993 were calculated from the half-hourly averages of upwelling long-wave (3–50 μm) radiation data. The site conditions on those days are given in Table 2. The rate of change of surface temperature from one period to the next was then calculated as degrees centigrade per hour. The rates are plotted individually by time periods in Figure 3 for groupings by type of ground cover and compared for

Table 2. Site conditions at SOROIDS. The range of wind gusts and air temperature for each day are based on maxium wind speed and 30-min-averaged air temperature, both at 2 m height, as recorded each half hour. The low insolation on 2 May and 13 January is due to overcast sky conditions associated with rainfall and snowfall, respectively.

				Insolation[a]		
Date	*Ground cover*	*Wind gusts (m/s)*	*Air temperature (°C)*	*Maximum (W/m²)*	*Sum (W/m²)*	*Snow depth[b] (cm)*
1992						
23 Feb	snow	0.2 to 7.2	–5.5 to 3.2	599	4,410	4
24 Feb	snow	0.2 to 3.2	–7.1 to 1.5	390	4,727	4
25 Feb	snow	1.8 to 9.8	–1.2 to 0.6	99	713	4→6
26 Feb	snow	0.2 to 7.8	–1.2 to 3.8	292	2,415	6
9 Apr	g-t-s[c]	0.4 to 10.8	–3.6 to 13.3	800	9,959	
10 Apr	g-t-s	0.8 to 10.8	1.0 to 13.7	702	9,006	
12 Apr	snow	1.2 to 13.9	0.1 to 7.7	520	5,227	5→3.5
14 Apr	g-t-s	0.6 to 13.0	–7.9 to 7.9	930	11,404	
15 Apr	g-t-s	0.5 to 11.3	–5.5 to 7.0	862	13,576	
29 Apr	g-t-s	0.2 to 11.9	–1.3 to 18.5	889	12,923	
30 Apr	g-t-s	0.5 to 8.6	0.2 to 18.0	815	7,742	
1 May	g-t-s	0.7 to 9.2	8.3 to 16.7	837	11,328	
2 May	g-t-s	0.6 to 13.0	7.9 to 16.0	424	4,584	
21 Dec	grass	0.9 to 13.9	–7.2 to –1.9	339	3,228	
22 Dec	grass	0.7 to 8.2	–4.5 to 4.2	334	2,902	
1993						
10 Jan	grass	1.2 to 4.0	–15.0 to –6.8	328	2,785	
13 Jan	snow	0.5 to 8.9	–5.6 to –1.2	131	604	0→16
2 Mar	snow	0.7 to 6.9	–17.8 to 4.9	553	5,498	44
16 Mar	snow	0.6 to 12.0	–18.3 to 7.5	693	8,091	62→56
23 Mar	snow	0.3 to 4.0	–11.3 to 7.9	752	9,187	46

a Maximum = Highest incident solar radiation (30-min average) recorded each day.
Sum = Daily total of averaged insolation values recorded every half hour.
b At the location of the acoustic snow depth sensor.
c g-t-s = grass–thatch–soil



a. Winter grass cover.

Figure 3. Time series record of rate of change of apparent surface temperature.

differences characteristic of the seasonally variable ground cover and associated weather conditions. The selection of days was based on: 1) nonwinter days with no rainfall, 2) absence of interfering activity at the site, and 3) completeness of the site characterization data. The exception to the first criterion is 2 May 1992: 4.2 mm of rain fell between 0100 and 0500 hours and 15.1 mm fell between 1700 and 2230, with intermittent light rainfall during the day.

Grass-covered ground

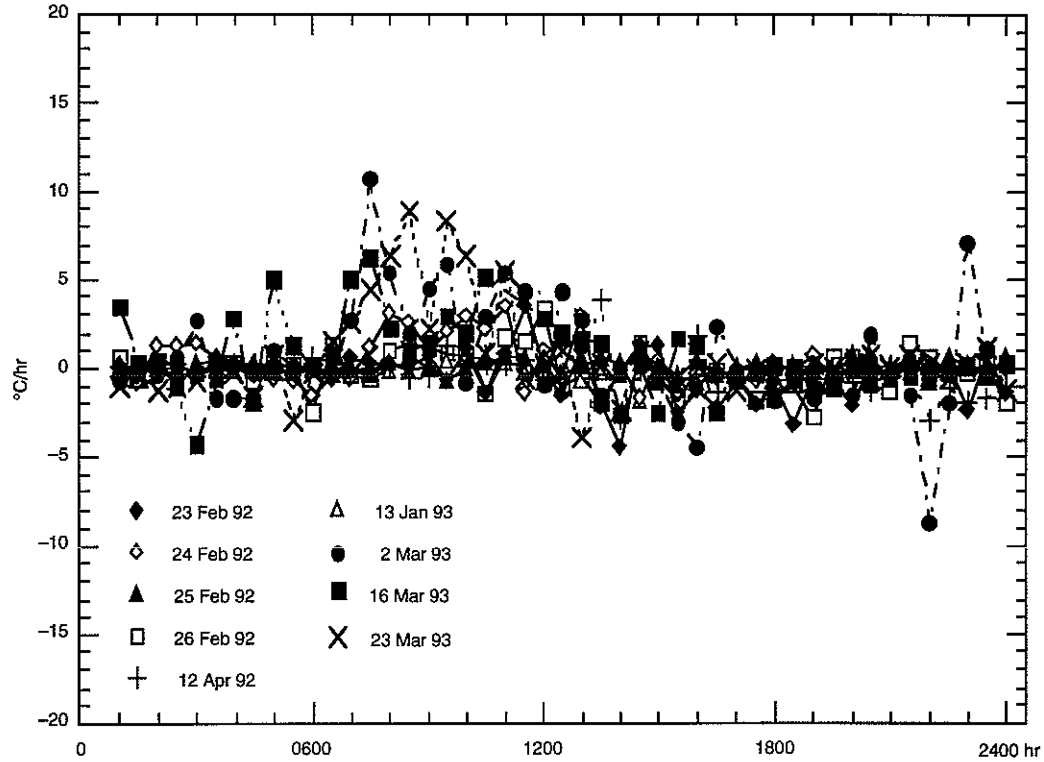
The rates of surface heating of a fully established, dormant grass cover in winter are represented by December and January days on which there was no snow cover (Fig. 3a). There is a clearly evident diurnal cycle to the rates of temperature change of the grass surface during the course of a day. The grass warms (positive rates) throughout the morning and cools (negative rates) in the afternoon. The maximum rate of change of surface temperature of the winter grass, 5°C/hr, is half that of a summer grass cover, 10°C/hr (Peck 1993). This can be directly attributed to the seasonal differences in weather conditions. The maximum surface temperature of the SOROIDS grass cover was 30–45°C in June and 16–25°C in September and October, but only −1 to 7°C for the December and January days with no snow cover that are listed in Table 2.

Snow-covered ground

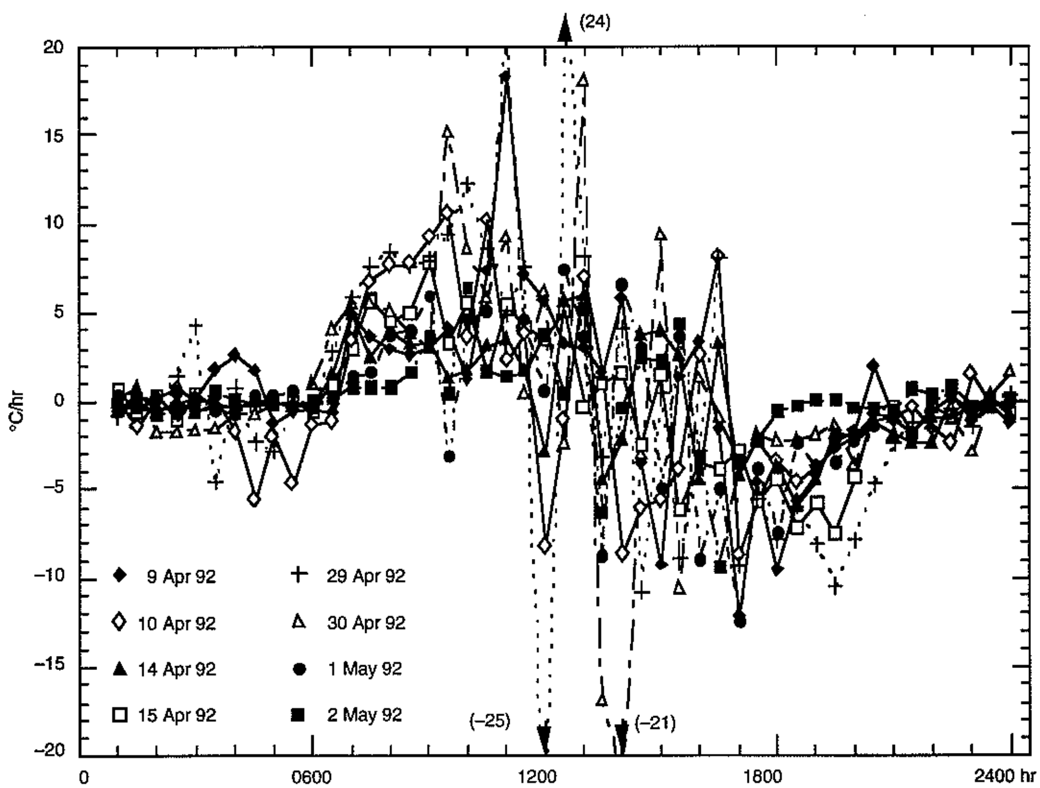
The rates of change of the surface temperature of a snow cover are given in Figure 3b. On some days (25, 26 February 1992) the rates of surface temperature change are predominantly less than 2°C/hr (absolute value) and vary little through the course of the day. These were days of moderate warming when the range of air and snow surface temperatures was less than 5°C. Other days (2, 16, 23 March 1993) show relatively high rates of change during the morning hours. On these days there was significant warming of the site during the morning, with air temperatures rising by 23, 24, and 19°C, respectively, during the day. The snow surfaces warmed sufficiently for melting to occur on all the days, thus preventing large temperature changes during the early afternoon. The high rates of temperature change on the March days are the consequence of higher insolation and extended daylight hours. Thus, there is a time-of-year variation to the stability of a snow cover as a thermal background, with a late-winter snow cover potentially being more dynamic than a midwinter snow cover due to the greater likelihood for significant warming during the course of a day.

Transitional ground covers

The April and May days selected for this category of ground cover span a much shorter time because the



b. Snow cover.



c. Grass–thatch–soil cover.

Figure 3 (cont'd). Time series record of rate of change of apparent surface temperature.

vegetated surface transitions rapidly from winter dormancy to summer lushness. The apparent surface temperatures used to calculate the rates strictly apply only to the distribution of thatch and exposed soil through which new-growth grass is appearing at the pyrgeometer location. The proportion of grass increases over the month represented.

There is a larger range of rates during daylight hours for this ground cover type, generally 20°C/hr maximum but occasionally higher (Fig. 3c). One reason for this is that the late winter period at SOROIDS includes many overcast mornings. This delays or reduces the morning rate of solar warming of the surface and often results in continued warming (positive rates) into the afternoon. A second reason is that, frequently, the ground surface radiatively cools to below 0°C at night, with dew or frost forming that must evaporate or melt before solar warming of the surface can proceed. A third reason is that the increase in surface temperature from nighttime low to daytime high is larger for the transitional grass–thatch–soil ground cover (29–45°C) than it is for the winter grass cover (9–13°C) or the snow cover (6–12°C midwinter and 22–25°C late winter). The large rate of radiative heat loss from the surface components (grass, thatch, soil) during the afternoon and early evening is followed by relatively stable surface temperatures through midnight. Unless radiative cooling is moderated by a cloud cover or by the introduction of warm air by a frontal passage or other source, the lowest surface temperature will occur shortly before sunrise (Rosenberg 1974).

On the one rainy day included in the data set (2 May), the rate of change of surface temperature of the grass–thatch–soil ground cover was low (maximum 6°C/hr) throughout the day. In the morning and early afternoon this was because of low radiant-energy heating during overcast periods following or preceding rainfall, respectively. At midday, the ground cover remained relatively cool because solar energy was consumed in evaporating rainwater from the wet grass and soil. The latent heat associated with the phase change was an energy sink that moderated heating of the surface.

EFFECTS OF GROUND COVER VARIATION ON PIR RESPONSE

The two thermal parameters that determine PIR response—the magnitude and the rate of change of thermal radiance—both significantly depend on the winter ground cover at SOROIDS, as the above examples have shown. The diurnal range of surface temperature and the rates of change of surface temperature are distinctly different for dormant grass, snow, and grass–thatch–soil ground covers.

The thermal radiance emitted by the ground cover surface in the wavelength band to which the PIR's thermal detectors are sensitive (8–14 mm), increases proportionally to the fifth power of the surface's temperature,

$$N_{8-14\ \mu m\ \mathrm{radiance}} = 1.876 \times 10^{-11} T^{5.032},$$

where T is degrees Kelvin. This equation was obtained from Planck's equation for the specific spectral band of 8–14 µm; it is valid over the temperature range of –50 to 50°C and for a surface with emissivity of 1 (Lacombe and Peck 1992). For the range of surface temperatures encountered at SOROIDS, the corresponding changes in radiance that can be ascribed to a 1°C rise in temperature are given in Table 3. Clearly, the warmer the background scene is, the more significant to the PIR is a 1°C rise in temperature.

The 30-minute-average PIR voltage tends to show little variation over 24 hours regardless of the ground cover; essentially, it represents the base instrument response associated with a lack of differential change in thermal radiance. The maximum and minimum voltages each half hour represent PIR response to the greatest difference (decrease or increase, respectively) in thermal radiance incident simultaneously at a PIR's two thermal detectors. If there were no spatial variability in thermal radiance from the ground cover, then average, maximum, and minimum voltages would be identical.

Grass cover on 10 January 1993

Only PIR-West was operating on this day (Fig. 4). There is a diurnal cycle evident in the PIR response, with the onset of pronounced changes in the maximum and minimum voltages in the morning coinciding with initial radiant-energy heating of the grass cover. Voltages approach their predawn levels during the afternoon

Table 3. Change in thermal radiance attributable to a 1°C change in temperature.

Temp. (°C)	*Radiance (W/m^2 sr)*	*Temp. (°C)*	*Radiance (W/m^2 sr)*	*Δ radiance (W/m^2 sr)*
–30	19.01	–29	19.41	0.40
–20	23.29	–19	23.75	0.46
–10	28.30	–9	28.84	0.54
0	34.14	1	34.78	0.64
10	40.91	11	41.64	0.73
20	48.72	21	49.56	0.84
30	57.68	31	58.64	0.96
40	67.91	41	69.01	1.10

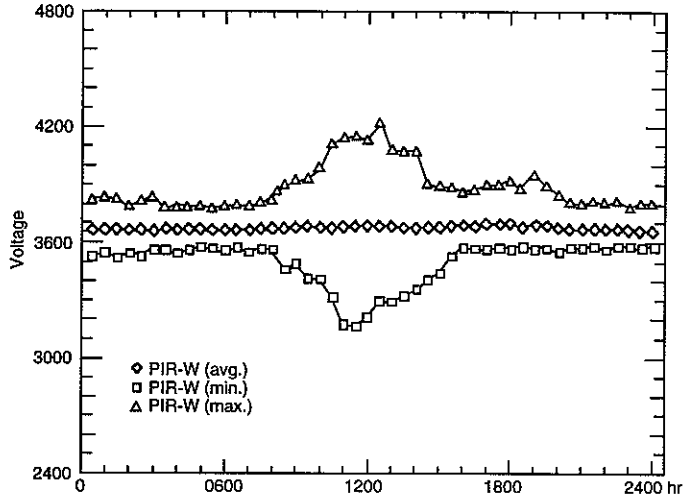


Figure 4. Time series record of PIR-West voltages on 10 January 1993 when the ground cover was dormant grass. Positive voltage change relative to the average indicates radiance change from high to low; negative, from low to high.

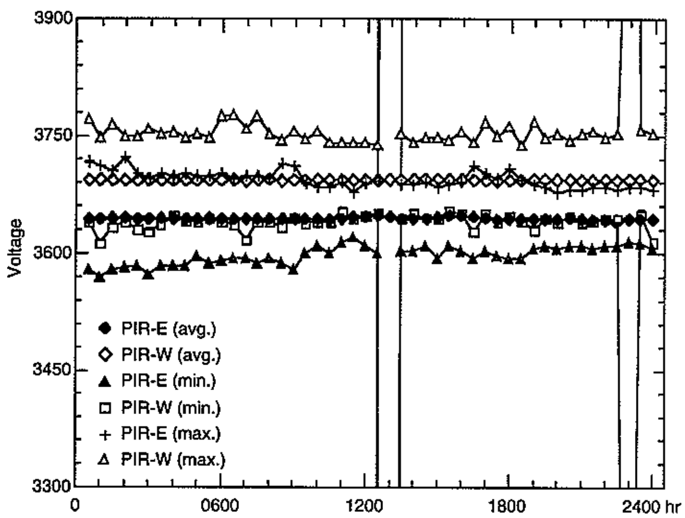


Figure 5. Time series record of PIR-East and PIR-West voltages on 24 February 1992 when the ground cover was a 2–4-cm-deep continuous snow cover.

as the grass radiatively cools. At night, when the surface temperature is unchanging (Fig. 2a), the PIR's voltages are steady.

Snow cover on 24 February 1992

The response of both PIRs on 24 February 1992 is shown in Figure 5. During daylight hours, first PIR-East, then PIR-West views the snow cover roughly in line with the direction of solar illumination. The off-scale voltages of both PIRs in the half-hour period ending at 1300 hours and for PIR-West at 2235 hours are responses to a person and an animal, respectively, crossing the PIRs' zone. The voltage range (maximum–minimum voltage for the same half-hour period) of both PIRs decreases overall from 0030 hours through 1030 hours. This corresponds to warming of the snow surface from –10 to –2°C; for the remainder of 24 February the apparent temperature of the snow surface (calculated from the upwelling [3–50 µm] radiation) was between –2 and 0°C. The PIR response is lowest (voltages closest to the average value) while the temperature of the snow surface is stabilized. The small variation in surface temperature after 1030 hours is evident in the low rates of temperature change, not exceeding 3.5°C/hr, that predominate on that day. This was a calm day with wind gusts less than 4 m/s throughout. The responses of the PIRs on other winter days when there is a snow cover at least several centimeters deep show similar low magnitude and small range of values even on windy days (gusts >4 m/s). A midwinter snow cover is consistently the most stable thermal background scene.

Grass–thatch–soil ground cover on 10 April 1992

The response of the PIRs on this transitional day is shown in Figure 6. The nighttime hours are characterized by low response (voltage levels close to average values) and low variation except for PIR responses to animals (half-hour periods ending 0430, 0500, 2400). The daylight hours are a period of persistently high voltage levels; this corresponds to first the solar warming and then the radiative cooling of the mixed grass–thatch–soil ground cover (Fig. 2c). Rates of change of the surface temperature were as high as 10.5°C/hr, three times the maximum rate of the 24 February snow cover and twice the rate of the 10 January winter grass. The high rates of temperature change together with the larger radiance from the warmer surface (Table 3) are the reason PIR response is highest for the grass–thatch–soil ground cover during daylight hours. Compared with those of the dormant grass cover and the snow cover, the surface temperature of the spatially diverse grass–thatch–soil ground cover changes nonuniformly within the fields of view of a PIR's two thermal detectors.

PIR RESPONSE TO A DISAPPEARING SNOW COVER

The examples above have shown that a time series record of PIR voltage is a useful indicator of the relative magnitudes of diurnal variations in surface temper-

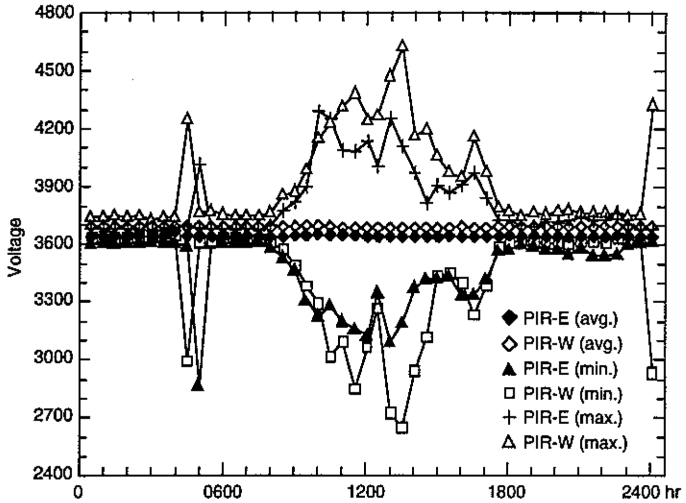


Figure 6. Time series record of PIR-East and PIR-West voltages on 10 April 1992 when the ground cover was a spatially heterogeneous mixture of new-growth grass, thatch, and soil.

ature, and the rate of change of surface temperature, as they occur differentially in the fields of view of a PIR's two thermal detectors. PIR response to a snow-covered background is low in magnitude and varies little during a diurnal cycle because of the low thermal radiance and spatial homogeneity (relative to the fields of view of a PIR's thermal detectors) of a continuous snow cover. A typical late-winter transitional ground cover of spatially heterogeneous grass–thatch–soil is a thermally dynamic background because the components heat and cool differentially as determined by their solar albedos and heat capacities, because of the extreme variations in surface temperature between daytime and nighttime, and because of the higher emitted radiances associated with the higher surface temperatures. PIR response to such a background is low and steady during nighttime or overcast periods when there is no solar heating of the surface, but it is high and erratic during sunny daylight hours.

Given that there are clear differences in PIR response to a continuous snow cover (e.g., 2–4-cm-deep snow cover of 24 Feb) and to the mixed grass–thatch–soil ground cover (e.g., 10 April) that underlies a late-winter snow cover, is it possible to determine from a PIR's response when a melting, late-winter snow cover no longer thermally shields the underlying ground cover? It is necessary to consider PIR response to a disappearing snow cover on a sunny day, during daylight hours, to have the largest potential contrast between the stabilized surface temperature of the melting snow and the variable (insolation-dependent) surface temperatures of the grass–thatch–soil ground cover. A hindrance to large increases in the surface temperature of the ground cover upon the disappearance of the snow cover is that the grass, soil, and thatch will all have been wetted by meltwater, so solar energy will be consumed in first drying the materials (latent heat of evaporation) and then in sensible heating. Despite having a database of PIR voltages that includes the late-winter periods of three years, there was only one occurrence when a snow cover disappeared rapidly during daylight hours. Commonly, either the snow covers melted at night, when there is no solar heating and so PIR response is always low, regardless of the surface type, or the shallow snow covers diminished more slowly, persisting in reduced depth over several days and becoming discontinuous in coverage, so the snow cover/no snow cover conditions were less distinct.

The one useful example of the loss of snow cover occurred on 18 April 1992. A storm on 16–17 April had deposited 5 cm of snow onto previously bare ground (grass–thatch–soil) between 2300 hours and 0630 hours. Snow depth decreased gradually following the cessation of snowfall on 17 April, remained at 3 cm overnight, and on 18 April decreased rapidly from 0700 to 1630 hours, when no snow remained in the area sampled by the acoustic snow depth sensor.

The response of both PIRs to this changing background scene is shown in Figure 7. The off-scale voltages of PIR-West for 2130 to 2300 hours are in response to the thermal radiance of small animals crossing in front of that unit. Compared with 24 February when there was a 2–4-cm-deep persistent snow cover, the voltages of both PIRs are higher and more erratic during the late morning and afternoon hours.

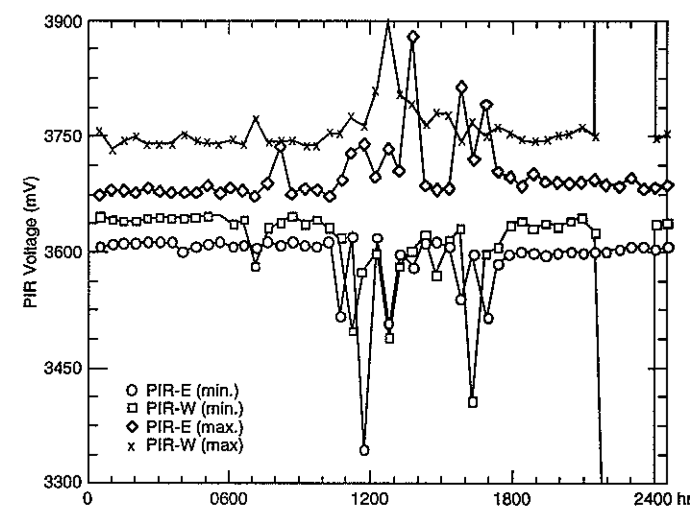


Figure 7. Time series record of PIR-East and PIR-West voltages on 18 April 1992. The 3-cm-deep snow cover melted completely during the day to reveal grass–thatch–soil ground cover.

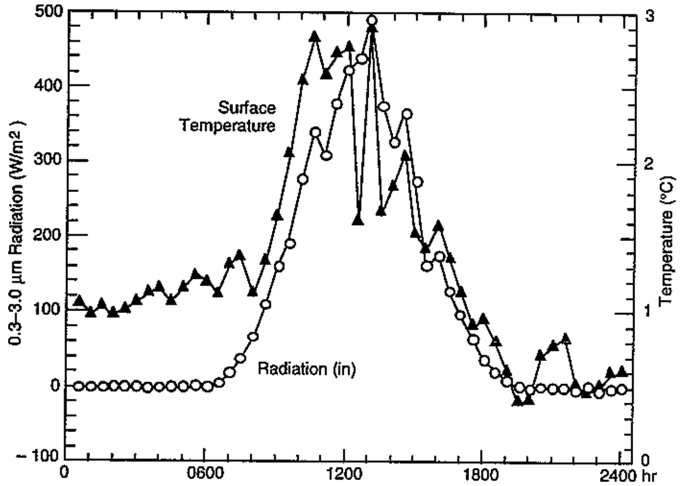


Figure 8. Time series record of ground surface (calculated) temperature and incident solar radiation (30-min average) on 18 April 1992.

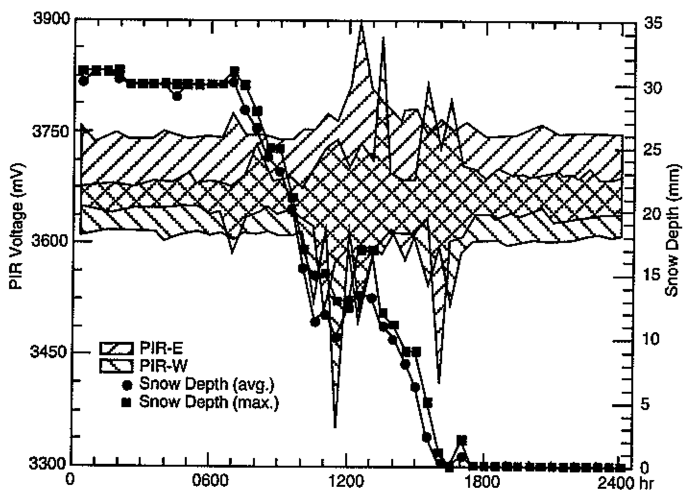


Figure 9. Time series record of PIR-East and PIR-West voltages and snow depth (maximum and average) on 18 April 1992. Each PIR's voltage range is shown as a shaded band.

Compared with 10 April, when there was no snow cover and strong insolation, the PIR voltages during daytime are quite low. Clearly, the spatial and temporal variation of the thermal radiance of the 18 April ground covers fits neither the winter snow cover nor the transitional grass–thatch–soil ground cover category.

Figure 8 shows the time series records of insolation and apparent surface temperature on 18 April. Based on the net radiance from the area sampled by the pyrgeometer, the surface underwent a rapid increase in temperature from ~1 to 3°C in response to increasing solar radiation. The 1°C surface is a continuous snow cover, as indicated by a video record of the site. [The greater than 0°C apparent surface temperature is a consequence of the accuracies of the pyrgeometer and of the equation used to convert emitted radiation to surface temperature.] The occurrence of higher PIR response (voltages strongly deviating from average) on 18 April is delayed relative to the onset of solar warming of the scene, and it is delayed relative to the onset of rapid surface warming. Instead, the responses of the PIRs change in magnitude when the snow cover becomes less than ~2 cm deep (Fig. 9), and continue to be erratic through dusk, when strong solar warming ceases.

This exercise indicates that a continuous snow cover of ~2 cm depth is sufficient to provide a thermally uniform background for the PIRs when the underlying ground cover is a mixture of thatch, soil, and short grass. This means that spatially (on the order of the size of the two sectors of the ground cover surface viewed by a PIR's thermal detectors) and temporally (the rate of change of thermal radiance from each detector's zone), the variation in thermal radiance from the background is slightly greater than that associated with a deeper snow cover, but much less than that associated with a bare ground cover of grass, thatch, and soil. Increases in surface temperature as the snow cover disappears are suppressed by the conversion of absorbed radiant energy first to latent heat of melting (snow) and then to latent heat of vaporization (drying of the wet grass, thatch, and soil).

A sensor system such as a PIR might be used to monitor when a continuous snow cover has become established on formerly bare ground, or when a melting snow cover is nearing disappearance. The potential resolution would be snow depth on the order of a few centimeters. Neither snow fall nor snow disappearance could be determined during nighttime because the variation in thermal radiance from a vegetated surface when there is no solar warming is too similar to that from a snow cover.

This exercise also indicates that a snow cover less than 1 cm deep is unlikely to result in the stable PIR response associated with a deeper, midwinter snow cover. For PIRs used as intrusion detection devices, this implies that the presence of a very shallow snow cover—a dusting to ~1 cm deep—will not eliminate, only reduce in magnitude, the diurnal variation in PIR response associated with grass or grass–thatch–soil ground covers. This effect persists after the disappearance of the snow, until the materials (grass, thatch, soil) wetted by meltwater have dried.

CONCLUSIONS

Winter ground covers (dormant grass, snow, grass–thatch–soil) and associated weather conditions produce distinctly different thermal background scenes. Characterizing ground covers by the diurnal variations of their surface temperatures and rates of change of surface temperature, and by their visual uniformity, provides an indication of the temporal and spatial variability of their thermal radiance. A uniform snow cover is the least dynamic background; a uniform, dormant grass cover in winter (no snow cover) exhibits a moderate diurnal cycle of radiance changes; and a heterogeneous ground cover of grass, thatch, and soil (no snow cover) is most dynamic.

By monitoring the ground with a passive infrared sensor system that responds to differential changes in thermal radiance, the characteristic differences in thermal radiances from these ground covers could be used remotely to determine when snowfall onto bare ground or when complete snow melt has occurred. Other times, the PIR response would provide a measure of the relative similarity of spatial distribution and magnitude of surface temperatures within the two triangular sectors of ground viewed by a PIR's thermal detectors. This information would have application in formulating an energy budget for the ground cover that incorporates a realistic representation of the spatial distribution of surface temperature and associated heat fluxes.

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