

Chapter 15

Antarctic Climate and Soils

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Abstract Climate change is having a significant impact on the ecosystems of Antarctica and has been the focus of scientific research in the past few decades. Our understanding of how large-scale weather patterns have changed has increased, but there is still way to go before we know how the shift in climate influences temperatures at microscales. Antarctica has one of the most severe climate extremes on the planet; it is the windiest, coldest, and driest. Combined with its geographic isolation, observational studies of climate change have been hampered due to lack of sufficiently dense monitoring systems. The soil exchanges energy with the overlying atmosphere (boundary layer) which influences its temperature through sensible and latent heat fluxes, both are poorly understood in the terrestrial landscapes of Antarctica. It has been shown that at least at the height of biological activity during summer, shifts in weather patterns across the Dry Valleys can have profound impacts on temperature and availability of liquid water, with downstream consequences for the biota. Therefore the terrestrial ecosystem responds rapidly to the larger climate system on a seasonal timescale.

15.1 Climate of Antarctica

Evidence continues to accumulate that climate change is having a profound impact in the Earth's cold regions; the impact on surface temperature has been more severe in the Arctic. The stratospheric ozone depletion, however, has been the flagship of anthropogenic influence on the Antarctic continent, with the increase in temperatures not as dramatic as the Arctic. Long-term monitoring of sea-ice extent, glacier mass balance, ocean temperature and circulation, permafrost, and air temperature

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indicate that concerns about pronounced warming in polar regions raised more than two decades ago by general circulation modelling experiments (e.g. Manabe et al. 1991) were well founded. Commenting on scientific findings, the intergovernmental panel on climate change (IPCC) fourth assessment report (AR4) stated with high confidence ‘that natural systems related to snow, ice and frozen ground (including permafrost) are affected’. Examples include enlargement and increased numbers of glacial lakes, increasing ground instability in permafrost regions, and changes in some Arctic and Antarctic ecosystems (Pachauri and Reisinger 2007).

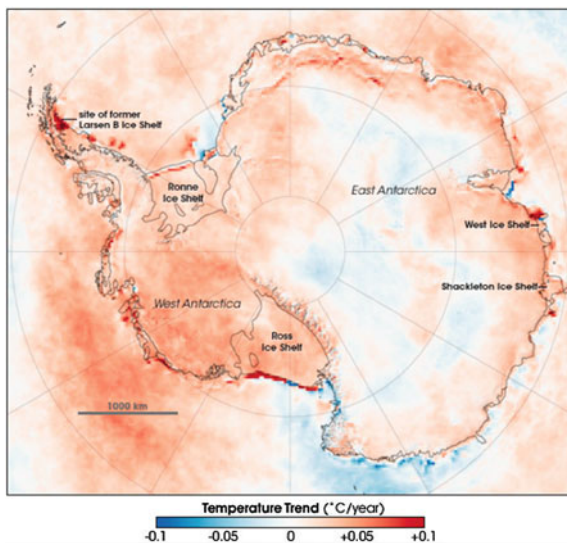
Antarctica has a vastly different climate regime when compared to the Arctic. Most opening paragraphs on Antarctic climate begin by stating that ‘Antarctica is the coldest, windiest and driest continent on Earth’. These characteristics are explained by the continent’s geographical location (which determines how much solar radiation it receives), the fact that it has the highest average elevation, and the effect of the uninterrupted Southern Ocean isolating the continent’s atmosphere. A comprehensive survey of knowledge on Antarctic climate was recently published by Scientific Committee on Antarctic Research (SCAR; Turner et al. 2009), covering not only the climate system in deep historical time (thousands of years into the past), but also the contemporary climate linkages with the global climate system (via tele-connections). This report was superseded lately to include new findings (Turner et al. 2013). Naturally, most climate research focuses on the spatial and temporal characteristics of either the snow/ice-covered regions of the continent, or the vast and dynamic sea-ice sheet that surrounds Antarctica, in this chapter; we will only examine the exposed soil regions.

Studies into climate (change) of Antarctica have been primarily hampered by paucity of in situ data. Despite the increase in the number of automatic weather stations (AWS) in the last 30 years, which stands at approximately 68 (Lazzara et al. 2012), spatial interpolation is often performed to map out geographical variation in unobserved areas. This is achieved by blending-in other data sets (such as remote sensing or numerical weather prediction data). Plans are in place to increase the amount of monitoring points, yet the harshness of the environment, and the logistical difficulties hamper efforts.

To extract climate signals, long-term records are needed, yet only a handful of AWS have operated for more than two decades; barely long enough to establish a significant trend (Schneider et al. 2011). When combining records to make a reconstruction, assumptions need to be made concerning selection criteria for the input data and the statistical methods to be used. Even with the generally large spatial coherence and correlation length scales of temperature anomalies at polar latitudes (e.g. Hansen et al. 1999; Chapman and Walsh 2007), none of the reconstruction methods can escape the basic limitation of few in situ observations in West Antarctica, and all exhibit less skill in this region compared with other regions of the continent, mainly due to the high variability in topography.

We have learned a great deal about the climate of the Antarctic since climate observations began. For the past 50 years, surface temperatures in Antarctica have increased by an average of 0.1 °C/decade (Schneider et al. 2011). But there is significant regional (West Antarctica vs. East Antarctica) variation and distinct

Fig. 15.1 Radiative surface temperature trends between for the period 1981–2007. (Source: <http://earthobservatory.nasa.gov/IOTD/view.php?id=8239>)

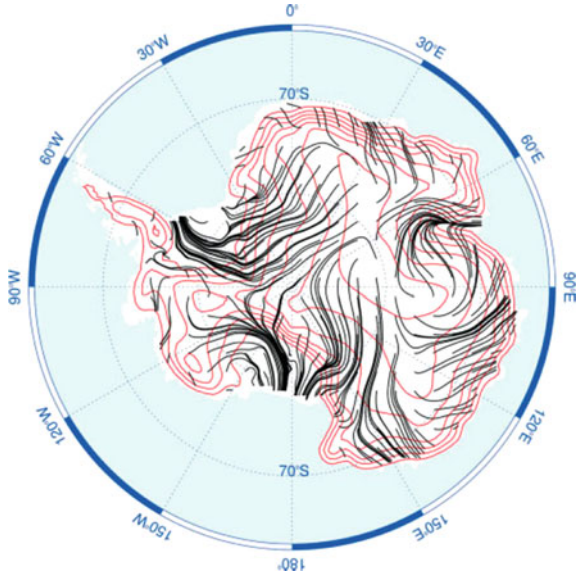


seasonality to the trends. East Antarctica has shown insignificant warming trends in summer and autumn (yet perhaps even some cooling), but with warming in spring and winter. In contrast, West Antarctica has experienced significant warming in spring (except for the Peninsula).

What is certain is that climate change signal varies across the continent. Some researches have recognized the enhanced warming in the Antarctic Peninsula (e.g. Rott et al. 1996; Chapman and Walsh 2007), and then on the other hand, Doran et al. (2002a, b) studied the McMurdo Dry Valleys for the period 1986–1999 and concluded that summer temperatures had decreased by 1.28 °C per decade. The autumn temperatures even showed a more dramatic decrease of more than 2.0 °C per decade. For the McMurdo Dry Valleys (MCM) in Victoria Land, Bertler et al. (2004) suggested that the short-term cooling was associated with El Niño-Southern Oscillation (ENSO)-driven changes in atmospheric circulation. In studying the effect of circulation patterns on Antarctic climate from 1957 to 2004, Marshall et al. (2007) reported that parts of the Antarctic continent are cooling. Considering the continent as a whole, AR4 states, ‘it is likely that there has been significant anthropogenic warming over the past 50 years averaged over each continent (except Antarctica)’ (Fig. 15.1).

The map in Fig. 15.1 is produced from thermal infrared (heat) observations made by satellite sensors. Satellites do not measure air temperatures, but can measure radiative surface temperatures by capturing the heat energy given off (emitted) by the surface. It shows that across most of the continent and the surrounding ocean, temperatures have climbed between 1981 and 2007. In some places, the rate of warming approached a tenth of a degree each year (2 °C over the entire period). Temperature rise was greater and more widespread in West Antarctica than in East Antarctica, where some areas showed little change or even cooled.

Fig. 15.2 Low-level katabatic wind patterns over the continent (adapted from Van Lipzig et al. (2004))

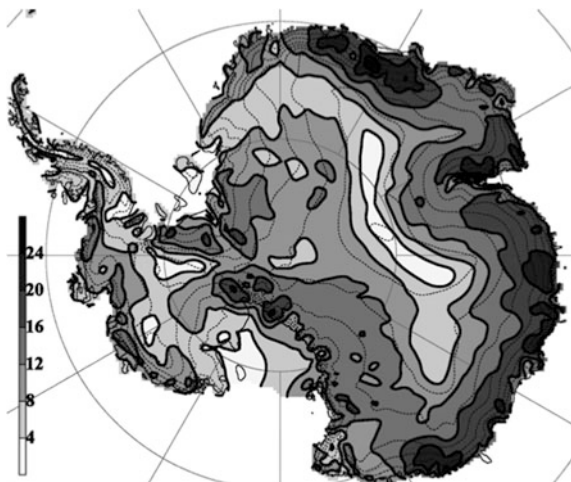


Antarctica is known for its katabatic winds, which influence the climate of almost the entire continent. Katabatic winds prevail when cold air is formed on a sloping terrain and simply drains towards lower elevation under the force of gravity. The ice/snow-covered sloped terrain produces katabatic winds almost constantly, directed by the topography, a diverging low-level flow pattern is set up (Fig. 15.2). The katabatic winds transport significant mass of air towards the continental margin and are in part responsible for the harsh weather experienced by early explorers.

The ‘windiness’ aspect of Antarctica is conveyed in Fig. 15.3. This map is produced by a computer weather model showing average annual wind speeds close to the ground. The coastal margins tend to record some of the highest sustained wind speeds recorded on Earth, because as the katabatic approaches the steep coastal topography, it interacts with the storm systems that circumnavigate the Southern Ocean and intensifies. The steeper slopes at the coastal margins, and the convergence of the katabatic through gaps (valleys) in the mountainous terrain also contribute to this intensification. The strength and variability of the katabatic winds can have significant impact on other systems, for example sea-ice. The increased intensity of the katabatic winds was shown to be responsible for the expansion of sea-ice (Holland and Kwok 2012) and is one of the factors in the sea-ice extent reaching its maximum recorded expansion in winter of 2012.

This chapter will focus on the state of knowledge regarding the boundary layer climates of the regions in Antarctica that have extensive areas of exposed soil. The boundary layer is the layer in direct contact with the surface. Exposed rock and soil total only about 46,000 km² of continental Antarctica (approximately 0.33 % of the land area) (Fox and Cooper 1994). The exposed soil exchanges energy with the

Fig. 15.3 Annual average wind speeds near the surface in meters per second (adapted from Parish and Bromwich 2007)



atmosphere, so both systems are coupled. To understand the heat flow in the soil, we need a closer examination of how energy is transferred between the two.

15.2 Boundary Layer Climate

Heat conduction into soils is important to understanding potential responses to changes in surface energy balance, particularly in the context of changing climate. Local energy exchange processes and transport of energy due to large-scale weather systems control the air and ground temperatures. During synoptically quiescent periods, the surface energy balance dominates how temperatures evolve in time, whereas in other synoptic conditions (such as storms), the transported air mass can have an overriding influence.

The radiation balance at the surface is defined as:

$$Q^* = (K \downarrow - K \uparrow) + (L \downarrow - L \uparrow)$$

where $K \downarrow$ down-welling short-wave solar radiation, $K \uparrow$ is the reflected amount by the surface, $L \downarrow$ is the down-welling long-wave radiation, and finally $L \uparrow$ is the long-wave radiation emitted by the surface. The balance of these components determines the net all-wave radiation balance (Q^*), which is basically the energy available for heating (if the balance is positive) or cooling (if the balance is negative) of the surface. Each of the individual radiative components can be controlled by a variety of complex factors; for example, $K \downarrow$ and $L \downarrow$ are controlled by cloud amount on short timescales, and by latitudinal location on seasonal timescales.

The microclimate is determined by the properties of the surface and how it handles the energy flow. The net all-wave balance is subsequently partitioned through the surface energy balance:

$$Q^* = Q_H + Q_E + Q_G$$

where Q_H is turbulent sensible heat flux, Q_E is the turbulent latent heat flux (both sensible and latent heat fluxes represent the exchange between the surface and the atmosphere) and Q_G is the subsurface conductive heat flux (this component can be either ice/snow or soil). Typically, if Q^* is positive and the surface has surplus of energy, the fluxes tend to flow away from the surface (i.e. the air above the ground gets heated, and if there is water available at the surface it will evaporate). For a more detailed explanation of boundary layer climates see Oke (1987).

The sensible and latent turbulent fluxes are complicated functions of atmospheric stability (state of turbulence) and vertical gradients of temperature and moisture in the atmosphere, but their descriptions are beyond the scope of this chapter. The conduction of energy in the ground is governed by Fourier's law defined as:

$$Q_G = -\lambda \left(\frac{dT}{dz} \right)$$

where Q_G at any depth is determined by λ (thermal conductivity of the soil), and the vertical gradient of temperature with depth (d = temperature difference between two levels). Thermal conductivity depends primarily on soil composition and water/ice content. The conduction of energy in the ground therefore is a function of surface energy balance, and energy (heat) can either propagate towards the surface or away from it. For the exposed ground in Antarctica, the ground layer can be broken into an active-layer, and a permafrost layer. The active-layer is underlain by permafrost and is subject to freezing and thawing cycles due to close coupling with fluctuations in atmospheric temperatures (Linell and Tedrow 1981) whereas the permafrost remains at or below 8 °C for more than two years (Linell and Tedrow 1981). The permafrost thermal regime reacts to climate variations at different timescales: (1) seasonally above the depth of zero annual amplitude (ZAA), (2) annually at the ZAA and (3) from years to millennia at progressively greater depths.

Air temperature and snow cover are tightly coupled to permafrost thermal regime and active-layer thickness through the surface energy balance (Guglielmin et al. 2003; Guglielmin 2004; Adlam et al. 2010; Zhang and Stammes 1998). Down-welling solar radiation ($K\downarrow$) can be especially important on bare ground surfaces (Adlam et al. 2010).

Naturally, most observational and modelling research into radiation and surface energy balances in Antarctica has focused on snow/ice-covered regions (King et al. 2006; Van den Broeke et al. 2006; Munneke et al. 2012).

In the MDV, surface energy balance has a stronger influence on the hydrological cycle than precipitation. The hydrological cycle, particularly melting, is

controlled by daily, seasonal and annual variation in surface energy balance (i.e. net all-wave radiation, sensible and latent turbulent heat fluxes) (Gooseff et al. 2011). Liquid water can be present at the surface due to surplus of net all-wave energy, which can drive sensible heat and warm local ambient temperature. Despite the 24-h solar radiation, glacier melt can still vary considerably on a diurnal cycle, due to solar elevation and topographic shading.

15.3 Soil Climate Monitoring

Recognizing the importance of monitoring ground temperatures, the scientific community established a network under the auspices of the global terrestrial network for permafrost (GTN-P). Ground temperature is monitored in boreholes in Antarctica with depths ranging from 2.4 to over 125 m. Temperature is monitored and logged at these sites within the active-layer and the permafrost to understand their long-term response under a changing climate. At some Circumarctic Active-Layer Monitoring—southern hemisphere (CALM-S) sites, soil temperature is measured at increments of 10 cm for the upper 1 m, with lower frequencies at depth. For a comprehensive description of the state of active-layer and permafrost monitoring see Vieira et al. (2010). Temporary networks are sometimes established for studies that need higher spatial resolution information. Adlam et al. (2010) established such a network for a 4-year period (1999–2003) in the McMurdo Sound region.

Although the permafrost layer's thermal state is important as a proxy for climate change, it is the active-layer that acts as a conduit for energy and controls heat conduction towards the permafrost layer (Fig. 15.4). The periodic warming pulses that the active-layer exhibits during winter, due to the air temperature pulsation that are controlled by warmer air intrusions from synoptic disturbances or katabatic winds, will propagate towards the permafrost layer and gradually over time increase its pre-summer energy storage making it more vulnerable to the continuous summertime solar radiation and thawing (Fig. 15.4).

Adlam et al. (2010) studied the soil climates in the McMurdo Sound region for a four-year period (1999–2003). The study investigated the active-layer depth for eight summers, but found the deepest active-layer depths occurred at all sites for the warm summer of 2002/2002 (Doran et al. 2002a, b). Therefore, the active-layer responds rapidly to interannual fluctuations in local climate. Yet there was no significant overall trend for the period of study. The duration of this research is probably too short for trend detection. Interestingly, they report that the mean summer and winter air temperatures, the total summer solar radiation, and mean summer wind speed explain 78 % of the variation in the active-layer depth (R^2 0.73).

Soil moisture can also contribute to soil heat conductivity, and although the soils in MDV are extremely dry, a considerable gradient can exist in summer along the many lakes and streams. A soil moisture gradient study adjacent to Lake Fryxell in Taylor Valley was presented by Ikard et al. (2009). They aimed to

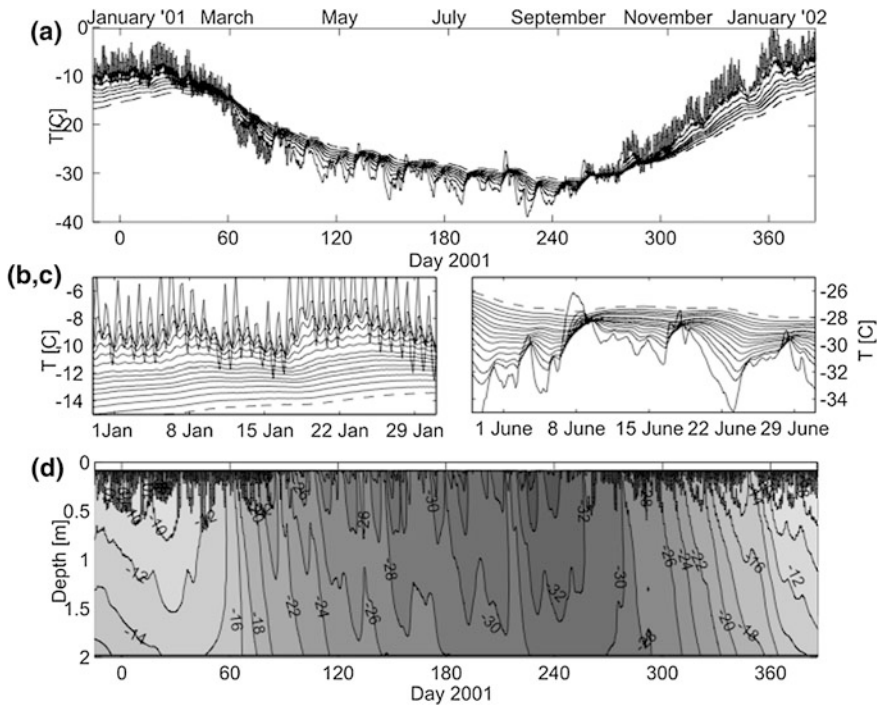


Fig. 15.4 Soil temperature time-series data near Ferrar Glacier in MDV. **a** The *bottom* temperature is at a depth of 1.98 m (*dashed line*) and the top temperature at 0.09 m has the largest amplitude (note: the figure is showing every second thermistor for clarity). **b** Reconstruction of soil temperature field; isotherm increments are 2 °C. From Pringle et al. (2003)

characterize the thermal characteristics of these relatively wet soils (as compared to the rest of the McMurdo Dry Valleys landscape) and to assess the response of the active-layer to possible increases in soil moisture. Estimated bulk ATD values were consistently greater in locations of enhanced soil moisture, so lakeside soils were more likely to conduct energy into the subsurface. Increased soil moisture across the landscape would likely increase ATD, allowing for greater heat exchange between the atmosphere and the subsurface.

Understanding the soil climate in the MDV can also aid in scientific fields such as astrobiology. The growing knowledge and understanding of surface processes on Mars have sparked an interest in the Dry Valleys of Antarctica as one of the closest natural terrestrial analogue of the Martian environment. Ideas and hypotheses can be tested there prior to launching probes into the distant planet, and careful analysis of corresponding periglacial terrestrial environments (Costard et al. 2002) is likely to aid in the interpretation of images gathered by Mars Orbital Imager (Malin and Edgett 2000).

15.4 Climate of the MDV

We consider in this section the state of knowledge regarding the boundary layer climate of the Dry Valleys. The ice/snow-free surface makes the energy exchange between the atmosphere and the ground significantly different from other locations in Antarctica. Although other ice/snow-free regions exist, MDV is by far the most extensive and studied region.

The MDV system is considered to be a polar desert, receiving little precipitation (annual water equivalent of 2–50 mm in the valleys bottoms; Fountain et al. 2010). Due to association with strong winds, precipitation is notoriously difficult to measure in MDV—and indeed in the rest of Antarctica. Little seasonal variability in snowfall is observed, but the spatial gradient can be quite pronounced, with values of 50 mm water equivalent along the coast sharply decreasing to 3 mm further inland (Fountain et al. 2010). The precipitation maximum at the coastal margin of Taylor Valley has been attributed to the winter convergence of local wind regimes that produce localized uplift. Snow patches accumulate as direct precipitation and drifting snow. Accumulation in valley bottom occurs in any wind-sheltered location, behind boulders, microtopographic features and within stream channels. Drifting snow by strong plateau wards wind intrusions, and glaciers account for roughly half of the measured snow accumulation.

Monaghan et al. (2005) investigated the reason(s) for the low precipitation amounts with a weather forecasting model. Using the one-year climatology generated by the model, they placed the MDV in the precipitation shadow region of the Transantarctic Mountains, where the moisture laden air streams from the large synoptic systems passing to the northeast and east of Ross Island do not penetrate the area. Most precipitation is in the form of snow and sublimates into the dry air prior to melting. But melting into ground is possible during summer with warmer ground temperatures and has been observed and modelled (Dallafior et al. 2012). Snow is a critical moisture source to the microbially dominated ecosystems in the soils of the Dry Valleys (Wall and Virginia 1999). Annual averages for air temperature hover near $-18\text{ }^{\circ}\text{C}$, while in winter they dip to $-50\text{ }^{\circ}\text{C}$ and for summer a few degrees below $0\text{ }^{\circ}\text{C}$ (Doran et al. 2002a, b).

With ample short-wave radiation, each valley in the MDV can develop its own distinct microclimate regime during the summer (Fig. 15.5). Characterizing the microclimate regimes in the MDV is the aim of New Zealand Terrestrial and Biodiversity Survey (nzTABS) and is achieved by AWS and shallow soil (2 cm) temperature measurements. There are marked variations in microclimates between the valleys; the soil layer can be above freezing point for a significant time in summer, and smaller valleys such as Miers can show significantly warmer monthly temperature (close to $5\text{ }^{\circ}\text{C}$ warmer). Miers is also a much more humid environment as reflected by the higher summer relative humidity.

In addition to the surface energy balance, horizontal transport (advection) of energy by winds is another factor in climate. In MDV, katabatic winds (drainage winds) are a characteristic feature. The larger-scale katabatic winds flow from the

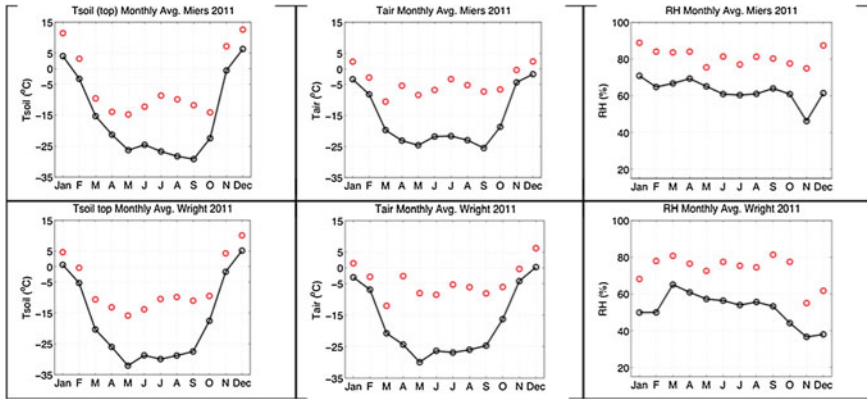


Fig. 15.5 Monthly averages (black dots) and the maximum daily value for the month (red dots) for Miers and Wright valley (data provided by NzTABS). RH is for air and not soil!

polar plateau at the western edge and can cause significant warming (particularly in winter) as they get adiabatically compressed when descending from higher elevations. They can reach significant intensity and flow above 20 m/s for prolonged periods. Air temperature increases by about 30 °C has been recorded over a few hours. This means that the frequency of these events has a significant impact on average temperatures; as it has been estimated by Nylen (2004) that for every 1 % increase in frequency, summer average air temperatures increase by 0.4 °C and winter average air temperatures increase by 1.0 °C (Fig. 15.6).

The interaction of boundary layer winds and the shallow soil temperature was examined for the Wright Valley by Zawar-Reza et al. (2012). This study looked at the variation of soil temperature with altitude from a series of temperature sensors placed along valley sidewalls for a winter season. The relationship between soil temperature and altitude shows that the valley bottom soils can be significantly cooler in winter—by as much as 15 °C, as cold air is trapped (ponded) by topography. Ridge-top areas are comparatively warmer, but this difference is occasionally obliterated as the strong katabatic winds warm the boundary layer within the valley and the soil.

15.5 Soil Climate and Global Change

The latest update on Antarctic's response to climate change states the likelihood of degradation in permafrost condition, and subsidence of ground surface amongst several other effects (Turner et al. 2013). As the atmospheric boundary layer warms and the surface receives more solar radiation, the active-layer is predicted to thicken. The thickening could have significant impact on terrestrial ecosystems. Areas deemed most susceptible are Antarctic Peninsula and coastal areas in East Antarctica.

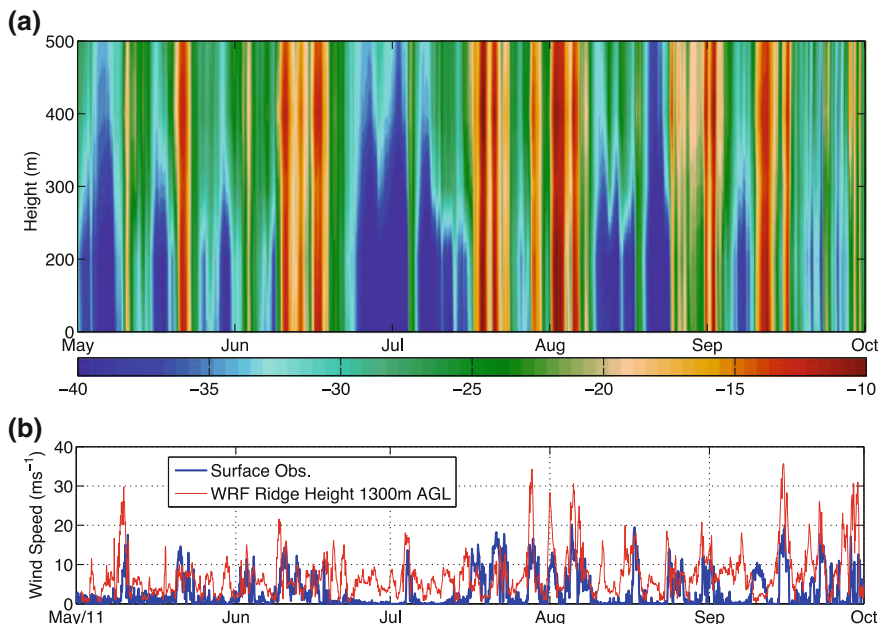
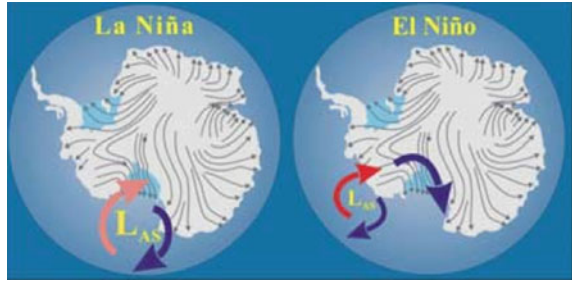


Fig. 15.6 **a** Variation of top soil temperature in winter for wright Valley, **b** linked with modeled and observed wind speeds

The Dry Valleys have cooled by $0.78\text{ }^{\circ}\text{C}$ per decade between 1986 and 2000 as determined from the in situ meteorological data (Doran et al. 2002a, b). This observation is contradictory to the concept of amplified warming in polar regions, and the substantial warming observed in Antarctica's Peninsula region over the second half of the twentieth century. The cooling reported by Doran et al. (2002a, b) is ecologically significant as it was more pronounced in summer (and autumn). The cooling trend is significantly correlated with decreased winds and increased clear-sky conditions, which can probably be attributed to more long-wave radiation loss from the surface. Since winter temperatures are well below the freezing point of water, winters do not have a significant direct impact on the hydrology or ecology. Therefore, summer temperatures are deemed to be critical drivers of the ecological system. The ecological response to the climate cooling included decreased primary productivity of lakes (6–9 % per year) and declining numbers of soil invertebrates (more than 10 % per year).

Later on, evidence was presented that linked the observed cooling to ENSO. ENSO is thought to be responsible for influencing the regional meteorology over the Ross Ice Shelf (Bertler et al. 2004). As a result, the authors challenged the idea that the cooling will be a long-term concern as ENSO exhibits variability on a decadal time scale. It appears that during La Nina events, the low-pressure system north of the Ross Ice Shelf is stronger and advects relatively warm and moist air from the north into the Ross Sea region. During El Nino, this flow is suppressed, as

Fig. 15.7 Climatological position of low pressure system in the Ross Sea. (From Bertler et al. (2004))



the low-pressure moves towards Amundsen Sea, forcing colder air masses through Siple Coast (Fig. 15.7).

What will be the specific effects of climate change on ecosystems in Antarctica? One study that addresses this question investigates how two unusual summers—one anomalously warm and the other unusually cold—affected the hydrology and the ecosystems in the MDV (Doran et al. 2008). The study found that changes in large-scale meteorological conditions had a major influence on the average temperatures. The synoptic features of the summer of 2000/2001 directed up-valley flows into MDV (air masses coming from the Ross Ice Shelf into the valleys), causing colder conditions, with an average seasonal temperature of $-6.1\text{ }^{\circ}\text{C}$. In contrast, when the synoptic drivers forced airflow down-valley, with air originating from the East Antarctic Ice Sheet in 2001/2002 season, the valleys recorded much warmer average seasonal temperature ($-3.7\text{ }^{\circ}\text{C}$). Since it was found that the difference in average short-wave radiation was negligible between the two seasons, the horizontal advection was deemed to have an overriding control.

The $2.4\text{ }^{\circ}\text{C}$ difference between the two seasons had a vivid impact, leading to significant glacier mass loss in Taylor Valley and 3,000–6,000-fold increase in streamflows, and also lake levels decreased slightly in the colder summer and increased between 0.54 and 1.01 m with warm conditions. As the authors noted, the warm season erased the prior 14 years of lake level lowering in just three months.

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