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Snow in the McMurdo Dry Valleys, Antarctica

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ABSTRACT: Snowfall was measured at 11 sites in the McMurdo Dry Valleys to determine its magnitude, its temporal changes, and spatial patterns. Annual values ranged from 3 to 50 mm water equivalent with the highest values nearest the coast and decreasing inland. A particularly strong spatial gradient exists in Taylor Valley, probably resulting from local uplift conditions at the coastal margin and valley topography that limits migration inland. More snow occurs in winter near the coast, whereas inland no seasonal pattern is discernable. This may be due, again, to local uplift conditions, which are common in winter. We find no influence of the distance to the sea ice edge. Katabatic winds play an important role in transporting snow to the valley bottoms and essentially double the precipitation. That much of the snow accumulation sublimates prior to making a hydrologic contribution underscores the notion that the McMurdo Dry Valleys are indeed an extreme polar desert. Copyright © 2009 Royal Meteorological Society

KEY WORDS Antarctica; Dry Valleys; snow; katabatic; polar desert; precipitation

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1. Introduction

The McMurdo Dry Valleys (MDV) are a polar desert with annual temperatures averaging −18 °C (Doran *et al*., 2002). Essentially, all precipitation is snow and can occur at any time of the year. Rainfall does occur occasionally, but only rarely reaches the valley floor and only then as a few drops (Keys, 1980). Despite roughly 40+ years of scientific investigation, little is known about the magnitude and spatial distribution of snowfall in the valleys. While snow on the valley floors has been correctly dismissed for being of little direct hydrologic importance (Chinn, 1981), this lack of interest had pervaded associated disciplines. We now realize that snow is a critical moisture source to the microbially dominated ecosystems in the soils of the Dry Valleys (Wall and Virginia, 1999). And snowfall patterns are important to the spatial patterning of dry and ice-cemented permafrost (Hagedorn *et al*., 2007). Coastal areas tend to have more snowfall than inland areas (Bull, 1966; Keys, 1980) where coastal areas have more ice-cemented permafrost than drier inland areas (Bockheim, 2002). Snowfall is probably critical to the survival of massive subsurface ice deposits. Some of these ice deposits may be of great age, $>10^6$ years (Sugden *et al*., 1995), yet how they are maintained over this time is unclear. Also, ice-cemented permafrost and massive ice are commonly observed close to the coast and are thought to be remnants from the last ice age >6000 years (Stuiver *et al*., 1981; Denton and Marchant, 2000). These massive ice deposits and other unknown deposits play a role in the current hydrology of the Dry Valleys when they melt during decadal warming events (Harris *et al*., 2007). While they contribute little water flux compared to that from the surrounding glaciers, they supply concentrated salts from the melting ice itself or dissolved from salt accumulations in the otherwise dry soils to the streams and lakes in the valleys affecting the biogeochemistry of the soils, streams, and lakes.

Snow is important to maintaining the glaciers in the valleys, and snow accumulation measurements are taken on several glaciers (Fountain *et al*., 2006). Summer snowfall on the glaciers is critically important to the hydrology of the valleys. Snow increases the albedo of the ice surface and changes the energy balance conditions such that little energy is available for melting (Fountain *et al*., 1999b). A snow accumulation of only a few centimeters largely eliminates glacial melt, and runoff from the glaciers slows to a trickle. To improve our understanding of the magnitude of snowfall and its spatial and temporal patterns, we report the results of our measurements from the past 12 years. Our study is focused on Taylor Valley, and we include results from several sites in two other valleys.

2. Site Description

The MDV is located on the coastal edge of the Antarctic continent (77.5 °S, 163 °E) in Southern Victoria Land

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(Figure 1). Only 2% of the Antarctic is ice free (Drewery *et al.*, 1982) and the MDV is the largest ice-free region on the continent with about 2000 km^2 of snow-free area (Chinn, 1988). The TransAntarctic Mountains block the ice flow from the East Antarctic Ice Sheet, which would otherwise cover the valleys. In places, outlet glaciers from the ice sheet flow into the valleys but terminate well before reaching the coast. In addition, numerous small alpine glaciers form in the mountains and flow to the valley floor. The valleys are characterized by a rocky–sandy soil devoid of vascular vegetation. The biology of the MDV is limited to microbial forms with spectacularly low biodiversity and short food chains (Wall and Virginia, 1999; Priscu *et al.*, 1999). Ephemeral streams, originating as glacial meltwater, supply perennially ice-covered lakes in the valleys (McKnight *et al.*, 1999). The polar climate of the region experiences continual darkness in midwinter and continual sunlight in midsummer. Air temperatures average about −17 °C with the winter minimum about −40 °C and the summer maximum a few degrees above freezing (Clow *et al.*, 1988; Doran *et al.*, 2002). In Wright Valley, annual values of snowfall have ranged 7–11 mm water equivalent (Keys, 1980; Bromley, 1985), most of which is lost to sublimation (Chinn, 1993). Katabatic winds are a typical feature of the regional climate, and commonly occur in winter and with less frequency in summer (Nylen *et al.*, 2004). These density-driven winds sweep down the polar plateau into the MDV with speeds up to 37 m s^{-1}, and their adiabatic warming can increase local air temperatures by 30 °C in less than a few hours. Snow can be transported, via the katabatic winds, from the Polar Plateau to the MDV (Keys, 1980; Bromley, 1985), acting like a precipitation event.

Little information exists on the snowfall of the MDV. The most comprehensive effort was conducted at Vanda Station in Wright Valley in the late 1960s and 1970s by the New Zealand Antarctic Program (Keys, 1980; Bromley, 1985). During this period, Vanda Station was continuously occupied for 3 years, providing a year-round record of the meteorological environment including precipitation. They recorded snowfall in every month of the year, except for one year when eight months had no snowfall (Bromley, 1985). The density of snowfall is quite low with a minimum value of 60 kg m^{-3} water equivalent (weq) and an average value of 100 kg m^{-3} (Keys, 1980). To our knowledge, since these earlier studies, no other precipitation measurements have been published until this present study. Spatial patterns of precipitation have been inferred from snow accumulation measurements on the glaciers (Fountain *et al.*, 1999a), and temporal patterns were inferred from snow pits (Witherow *et al.*, 2006; Bertler *et al.*, 2004). However, snow accumulation results from mass gain due to snowfall and mass loss from ablation; consequently, accumulation provides only a rough proxy for precipitation.

3. Methods

All precipitation gauges were deployed near pre-existing meteorological stations in the valleys to make use of the solid-state data loggers and to use the other meteorological variables in interpreting the precipitation record. The gauges were of three types: weighing bucket and

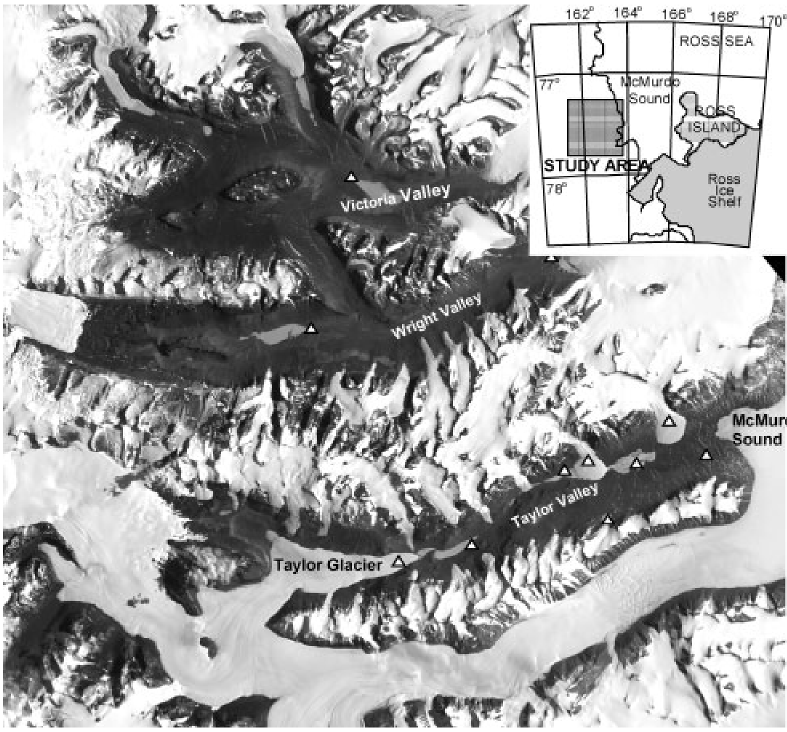

Figure 1. Landsat-7 satellite image of the McMurdo Dry Valleys. The image was acquired on December 19, 1999. The width of the image covers about 85 km. The triangles are locations of the meteorological stations.

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tipping bucket, both providing a water equivalent measure of snowfall, and a sonic ranger that provided a distance measure to the surface, yielding a snow depth. The weighing bucket gauges (Belfort Instruments, Baltimore, MD) consisted of a bucket on a digital weighing scale (manufacturer's accuracy 2.5 mm weq, precision of 0.5 mm) contained in a metal housing with an orifice about 1 m above the ground. A Nipher Shield (Goodison *et al.*, 1983; Yang *et al.*, 1999) was mounted over the orifice to increase gauge efficiency. The lip of the shield was about 1.6 m from the ground. The gauge housing and Nipher Shield were staked to the ground (Figure 2). As the precipitation accumulated in the bucket, the increased weight was recorded at hourly intervals. The bucket was initially partly filled with antifreeze to melt the captured snow, preventing it from being blown away. Also, a thin layer of silicone oil was added to the surface of the antifreeze to reduce evaporation. The contents of the bucket were emptied every 2–3 years and replaced with new antifreeze and silicone oil. We eventually replaced the weighing bucket at Lake Hoare station with a tipping bucket (Texas Electronics, Model: TE525MM); manufacturer's accuracy 0.1 mm weq. Snowfall is captured and melted in a reservoir containing an antifreeze mixture with a thin layer of silicone oil. The added mass overflows the reservoir into a tipping mechanism. The tipping gauge was placed within the housing of the weight gauge to make use of the Nipher Shield and the data recorded every 15 min.

Because of the expense and logistics required to purchase, install, and maintain the weighing bucket gauges, we deployed ultrasonic distance rangers (Campbell Scientific, Inc, model SR50; Judd Communications Ultrasonic Depth Sensor) to expand our spatial coverage. The rangers measure a distance to a surface using the travel time of an acoustic pulse. The rangers were installed on an arm extending from the station, 0.5–1 m above the surface (Figure 2). The manufacturers stated accuracy of the rangers was 10 mm and the distance was recorded hourly. To estimate water equivalent values, we episodically made manual measurements of snow depth and density.

The records for all gauges were examined to remove artifacts induced by instrument performance (see Results section). To minimize noise, owing to the low precipitation compared to instrumental sensitivity, we calculated daily averages for each site. If the difference between consecutive days exceeded a threshold of 0.5 mm weq for weighing buckets and 5 mm for the sonic ranger, the difference was added to a monthly sum. The thresholds are below the stated accuracy of the ranger because under most conditions they reliably detected smaller changes and our averages damped the noise (wind and electronic). To be sure that the noise was not inducing artifacts, the standard deviation of each daily average was calculated and, if unusually large, the record for that day was reexamined. No special handling of the tipping bucket data was required other than examination for artifacts.

4. Results

4.1. Instrument Performance

Three weighing buckets were initially deployed; however, early records, particularly in winter, were useless owing to the shaking of the instruments by katabatic winds. We attempted to secure the instruments with mixed success because of the lack of exposed bedrock. Finally, the gauge at Lake Bonney was sufficiently secured to provide good records. Yearly cleaning of the instrument was required to prevent the accumulation of fine sediment from interfering with the gauge operation. Once we achieved reliable data from Lake Bonney, we noticed two interfering factors. First, the more severe katabatic events transport sand into the buckets, mimicking precipitation events. These severe events

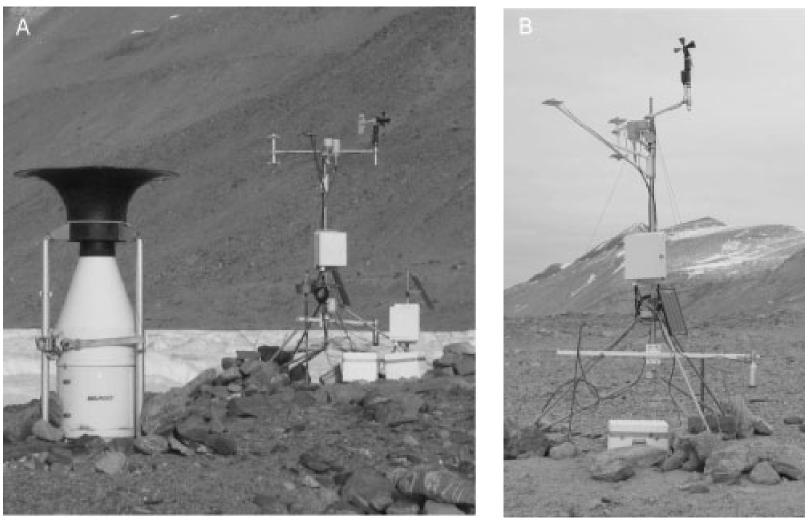

Figure 2. Photographs of meteorological stations in the McMurdo Dry Valleys. Photograph A is the station at Lake Bonney. The large black 'trumpet' is a Nipher Shield over a Belfort gauge. Photograph B is the station at Lake Fryxell with a sonic ranger suspended from a horizontal arm and pointing to the ground on the right side of the station.

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typically occur only a few times a winter. Second, diurnal changes in air temperature and solar heating of the gauge housing apparently caused expansion and contraction of the weighing mechanism, yielding apparent snow events. The apparent events were certainly small, $\leq$0.5 mm weq, occurring mostly in the summer months (November–January) when temperatures range from $-10\,^{\circ}$C to $4\,^{\circ}$C. In winter, apparent precipitation events coincided with katabatic winds that can warm the air by $30\,^{\circ}$C in a few hours (Nylen *et al*., 2004), yielding an apparent precipitation of 1.5 mm weq. However, unlike the real event that permanently increased the weight of the bucket, these events temporarily increased the apparent weight and returned to pre-event levels as the air cooled.

The tipping bucket gauge at Lake Hoare did not adequately record the timing of snow events because the light, low-density snow tended to collect in the gauge orifice bridging the cup of the tipping mechanism. With sufficient snow, the bridge was broken and the snow accumulated in the cup causing it to tip. As a result, we use the data to record snowfall amounts and not the exact (hours) timing of the snowfall. The tipping bucket was also sensitive to the high winds. Proper operation of this tipping bucket gauge required the reservoir to be filled to overflow capacity, making it quite sensitive to vibrations. Katabatic winds caused the antifreeze reservoir to overflow, tipping the bucket creating a false snowfall event and subsequently missing future events owing to a lowered reservoir until refilling by more snow. The gauge could never be secured sufficiently to eliminate the katabatic-induced vibrations, therefore the data recorded had to be handled carefully.

The ultrasonic rangers generally worked well and were reliable; however, the acoustic transducer could fail any time after 2 years and, until we established a routine replacement schedule, the record from the network of rangers was subject to numerous data gaps. The stated operational conditions of the rangers included temperatures down to $-40\,^{\circ}$C but the records became noisy at about $-35\,^{\circ}$C and were considered unreliable. Again, because we were measuring precipitation events that often resulted in thin accumulations of snow $\leq$10 mm, we were operating at the resolution of the instrument, whereas larger accumulations would have improved the signal-to-noise ratio. Fortunately, these cold temperatures commonly occurred during inversions (clear skies). Because surface detection of the ranger is based on the speed of an acoustic pulse, the signal also becomes noisy in high winds ($>$18 m s^{-1}) again obscuring the true signal (Brazenec and Doesken, 2005). On rare occasions, low-density snow events cause erroneous values and appear as spikes in the data. They are obvious and easily removed.

4.2. Data

The longest records of snow accumulation are Lake Bonney (12 years) and Explorers Cove (5 years), with most of the remaining sites collecting for 4 years (Table I). Annual snow accumulation in Taylor Valley did not exceed 100 mm water equivalent (weq) with minimum values as low as 3 mm. Close examination of the record shows that, in addition to precipitation, katabatic events may convey snow to the valley bottom, presumably from the snow-covered Antarctic Plateau to the west and from the local glaciers and snow-covered parts of the adjacent mountains. We presume that precipitation does not occur during katabatic events because the meteorological conditions that trigger such events are not favorable for precipitation. Therefore, we recognize two kinds of snow events, wind drift and direct precipitation, and we refer to the sum as 'accumulation' of which a fraction is 'precipitation'. We defined katabatic events following Nylen *et al*. (2004), having wind speeds $>$5 m s^{-1} and coming from the ice sheet (westerly). Snow accumulation during these events was considered wind drift. With katabatic events removed, annual precipitation at each site varied from 36 to 80% of the annual accumulation with an overall average of about 50%. The maximum value of annual snow accumulation was 98 mm weq in 2004 at the Explorers Cove site of which precipitation accounted for only 52 mm weq (53%). For the two sites with relatively long records, Explorers Cove (5 years), and Lake Bonney (12 years), the fraction of direct snowfall at each site was 50 and 43% respectively. The more severe katabatic events transport sediment as well, which was also captured by the weighing bucket gauges, as explained. Comparing the weighing bucket to the sonic ranger at Lake Bonney over the 2 years of overlapping data shows that, according to the bucket, 39–45% of the snow accumulation is precipitation, whereas the sonic ranger indicated 80–100%. We have no way of filtering the data to remove the effect of the sediment on the bucket weight, so the magnitude of snow accumulation at these two sites is probably overestimated and the fraction of that as precipitation is underestimated.

Monthly values of snow accumulation show little consistent seasonality at Lake Bonney, near the head of Taylor Valley (Figure 3). But at Explorers Cove, near the coast, snow is greatest in winter from March through September. In contrast, the nearby McMurdo Station ($\sim$60 km away) exhibits much more snow but no seasonality. For Taylor Valley, given the very low accumulation/precipitation values, any one storm can shift the seasonal or monthly balance of snow accumulation, and our records are too short to make any statistically significant statements of seasonal trends in snowfall. Longer-term trends over the 12 years of record at Lake Bonney (Figure 4) suggest increasing snow with time. However, the trend is not significant because the greatest snowfall occurs in the last 3 years of record, and because of the measurement uncertainty associated with the extremely low values.

A strong spatial gradient exists for Taylor and Wright Valleys with more snow closer to the coast and less snow inland (Figure 5). In Taylor Valley, the gradient of snow accumulation is -2.6 mm weq km^{-1} from the

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Table I. Accumulation data (mm water equivalent) in the McMurdo Dry Valleys. The first eight sites are in Taylor Valley; the first site is closest to the coast and the last site is further away. Lakes Brownworth and Vanda are in Wright Valley and Lake Vida is in Victoria Valley. Common. Gl stands for Commonwealth Glacier. G, gauge type; U, ultrasonic; W, weighing bucket; T, tipping bucket. The numbers in bold are accumulation values and italic is precipitation, that is, accumulation with snow drift from katabatic events removed. The specific location of the meteorological stations can be found in Doran *et al*. (2002) and Nylen *et al*. (2004).

Station	G	1995	1996	1997	1998	1999	2000	2001	2002	2003	2004	2005	2006
Explorers Cove	**W**						**42**	**66**	**89**	**66**	**98**		
							17	*24*	*47*	*42*	*52*		
Common. Gl	**U**									**43**	**47**	**39**	**93**
										36	*37*	*35*	*75*
L. Fryxell	**U**									**34**	**34**	**21**	**46**
										34	*32*	*20*	*37*
Howard Gl	**U**										**20**	**29**	**87**
											2	*28*	*74*
Canada Gl	**U**									**16**	**17**		**13**
										8	*14*		*13*
L. Hoare	**T**								**13**	**13**		**20**	**19**
									3	*6*		*2*	*19*
L. Bonney	**W**	**3**	**6**	**6**	**14**	**6**	**14**	**5**	**10**	**11**	**46**	**22**	**23**
		2	*3*	*3*	*6*	*1*	*3*	*4*	*8*	*5*	*18*	*8*	*10*
	U									**3**	**5**		
										3	*4*		
Taylor Gl	**U**									**6**	**6**	**13**	**14**
										6	*6*	*9*	*12*
L. Brownworth	**U**										**26**	**51**	
											13	*20*	
L. Vanda	**U**										**18**	**32**	
											10	*30*	
L. Vida	**U**										**35**	**22**	**45**
											33	*22*	*45*

coast to the Nussbaum Riegel, an 800-m-tall hill in the center of the valley. The gradient up valley from the Riegel is -0.3 mm weq km^{-1}, very similar to that in the adjacent Wright Valley (-0.4 mm weq km^{-1}). These gradients are nearly identical to snow accumulation measurements collected on the glaciers as part of the glacier mass balance program (Fountain *et al*., 2006). Snow accumulation on the glaciers between elevations of 200–300 m decreased at a rate of -2.6 mm weq km^{-1} from the coast to the Nussbaum Riegel; further up valley, the gradient decreased to -0.6 mm weq km^{-1}. In addition to an inland gradient, a north–south gradient may also exist, with more snow accumulation toward the north.

4.3. Persistence

In addition to snow accumulation and precipitation in the MDV, the persistence of snow on the valley floor is important for biological processes in soils as a source of moisture (Gooseff *et al*., 2003; Wall and Virginia, 1999) and light penetration into lakes (Howard-Williams *et al*., 1998). Snow is also a moisture source for maintenance of ice-cemented permafrost against sublimation losses (Hagedorn *et al*., 2007). The sonic rangers provide a measure of the snow cover persistence in the valleys. As one would expect, persistence is longer in winter than in summer. At Lake Vida, Victoria Valley, snow persists all winter, whereas it persists only a few days to a week at Lake Bonney in Taylor Valley (Figure 6). Of course, persistence is related to snowpack thickness and ablation rate (sublimation, melting, and wind erosion). Generally speaking, winter snowpacks are maintained for a month or so and summer snow lasts for a few days to a week at most.

5. Analysis

We compare the spatial characteristics of the precipitation measurements against those predicted by a numerical atmospheric model to determine whether our sparse measurements agree with theoretical expectations. The model fields are also used to infer the broader precipitation distribution across the MDV. Model results were derived from the Antarctic Mesoscale Prediction System (AMPS), an experimental weather prediction program that supports the daily operations of the US Antarctic Program (Powers *et al*., 2003). The AMPS simulations used for this analysis employed a version of the fifth-generation Pennsylvania State University/National Center for Atmospheric Research Mesoscale Model adapted for polar conditions – Polar MM5 (Bromwich *et al*., 2001 #6981). AMPS ingests observations from surface stations,

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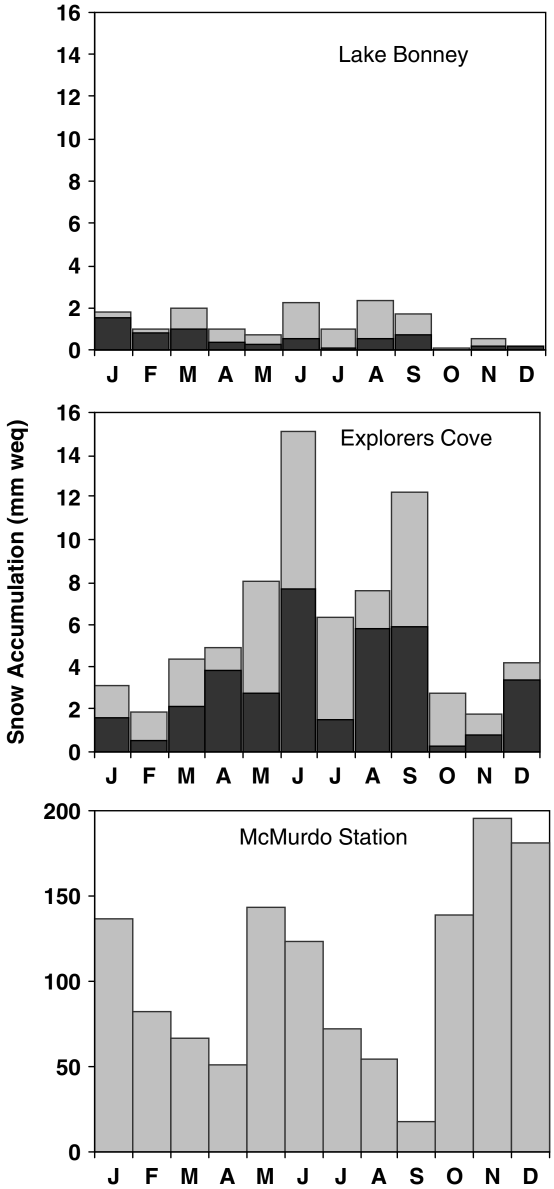

Figure 3. Average monthly snow accumulation and precipitation, Lake Bonney over the period, 1995–2006; Explorers Cove 2000–2004; and McMurdo Station 1973–1998. The gray bars are snow accumulation and the black bars are the precipitation (katabatic events removed). Snow accumulation at McMurdo Station is included for comparison; note the difference in scale.

radiosondes, satellite-derived cloud track winds, and sea ice data. Because AMPS has relatively high spatial resolution over the Ross Island vicinity where the hub of US logistical activities is, it is a useful tool for studying the regional climatology in detail (Monaghan *et al.*, 2005). Here we use 2.2-km spatial-resolution AMPS precipitation fields to assess MDV precipitation for a 1-year period spanning September 2006–August 2007. This period was chosen because it occurs shortly after AMPS was upgraded from 3.3-km to 2.2-km resolution over the MDV, which is sufficient to resolve the large topographic variation characteristic of the region. While employing more than one year of data would be ideal for a climatological study, constraints due to the size of the dataset limit us to a year. Monthly, seasonal, and annual precipitation fields were compiled from hourly AMPS output by assembling the 13- to 36-hour forecasts from the 0000 UTC AMPS simulation initialized on each day of the year. A previous comparison of AMPS results to observations for a similar model configuration indicates that the model fields are suitable for climatological applications (Monaghan *et al.*, 2005).

The model shows a precipitation shadow to the north of Mount Discovery (Figure 7), resulting from the precipitation-bearing synoptic systems that pass to the north and east of Ross Island (Simmonds *et al.*, 2003). The precipitation shadow diminishes to the north of Victoria Valley. Within the valleys, the precipitation increases with elevation and decreases from the coast inland. A local precipitation maxima at the mouth of Taylor Valley is probably due to the convergence of southwesterly katabatic flow during winter as it slows down upon reaching the mouth of Taylor Valley (Monaghan *et al.*, 2005). The uplift here may be further enhanced by southeasterly winds that commonly flow from the Ross Ice Shelf and into McMurdo Sound region. Unlike the valleys to the north (Wright, Victoria), Taylor Valley is open to the ocean without significant elevation to increase precipitation through forced uplift. Each valley has a precipitation minimum. For Victoria and Wright Valleys, the minima are associated with the lowest elevation landscape of the valley. The minimum in Taylor Valley is found inland of the Nussbaum Riegel, the 800-m-high hill in the center of the valley, which may contribute to a localized precipitation shadow on its upslope side (i.e. Figure 5). The model data shown in Figure 7 agree qualitatively with the spatial distribution of precipitation inferred from the ground measurements in the valleys. Modeled evaporation (not shown) is much larger than precipitation in the valley bottoms, consistent with the observations (Figure 6), which indicate that snow does not persist for long.

To more closely examine the details of the spatial changes, we compare the modeled precipitation with that measured in Taylor Valley where we have the highest

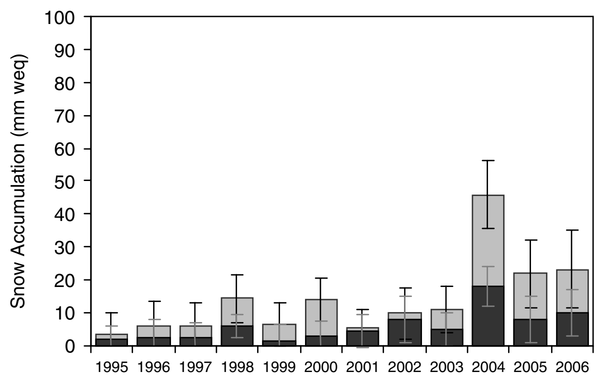

Figure 4. Snow accumulation at Lake Bonney. Gray bars represent the total snow accumulation for that year and black bars are the precipitation (katabatic events removed). Error bars are based on accuracy of instrument (2.5 mm) and the number of days with events recorded.

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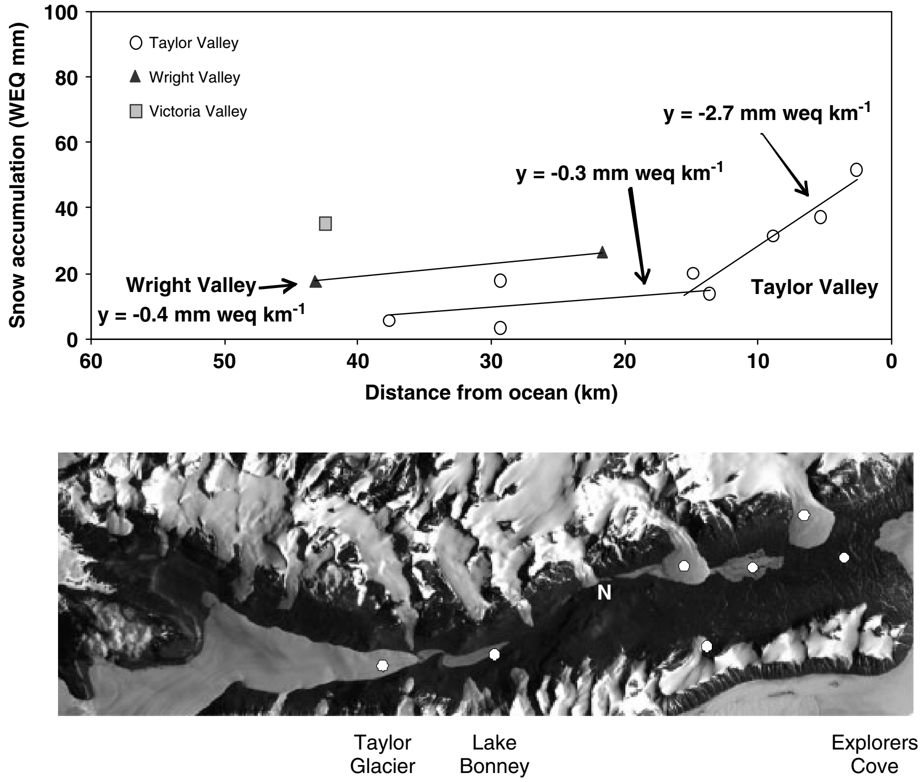

Figure 5. Spatial gradient of snow accumulation in Taylor and Wright Valleys for 2004. The Landsat image of Taylor Valley depicts the same horizontal scale as that of the graph. The white dots on the image show the locations of the meteorological stations and N indicates the location of the Nussbaum Riegel.

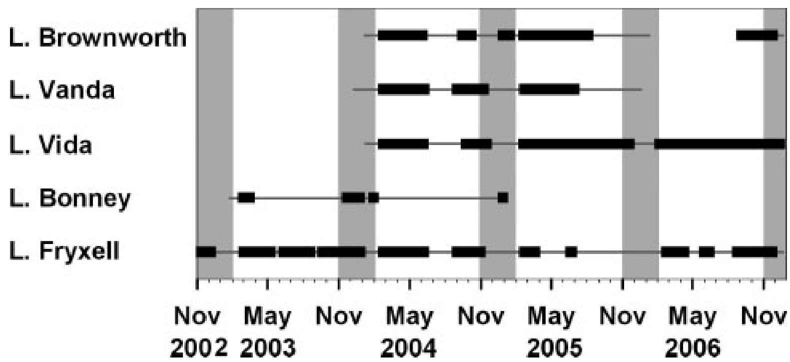

Figure 6. Duration of snow cover at all the stations with sonic rangers from November 2002 to November 2006. The vertical gray bars indicate the summer season from November through January. The period of record for each sonic ranger is indicated by a horizontal line and the boxes indicate the presence of snow thicknesses >10 mm for at least one day.

density of precipitation gauges. Results show that the modeled precipitation is greater than the observed mean precipitation (Figure 8). The overestimated precipitation is probably due in part to the model's topographic representation, which is somewhat smoother than the observed topography of Taylor Valley, especially near the Nussbaum Riegel. However, the observed error bars indicate that some of the difference is due to the limited model sample size (1 year); the maximum observed values overlap the model values at several stations. Despite the model bias, Figure 8 indicates that the model is able to capture the general trend of decreasing precipitation inland from the coast within the narrow walls of Taylor Valley.

6. Discussion

Given the extreme winter conditions and long unattended operation of the instruments, we found that using both weighing bucket and sonic gauges worked well together. Each instrument has its own operational issues but together they provided a fairly complete depiction of snowfall and accumulation. Annual total snow accumulations are low, as expected from previous studies (Keys, 1980; Bromley, 1985; Witherow *et al.*, 2006; Bull, 1966; Fountain *et al.*, 1999a; Fountain *et al.*, 2006) and are due to the precipitation shadow created by the TransAntarctic Mountains (Monaghan *et al.*, 2005). The largest annual accumulation, 98 mm weq, was measured

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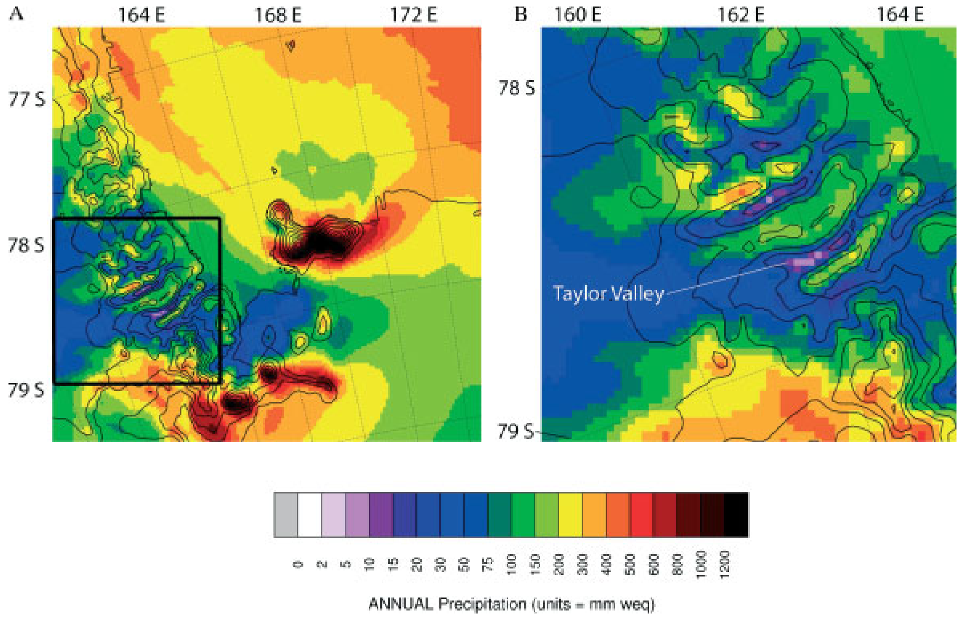

Figure 7. A: Simulated annual total precipitation for the McMurdo Sound region using the AMPS 2.2 km Polar MM5 (Powers *et al*., 2003). Ross Island is near the center of the image, and Mount Discovery is to the right of the lower right-hand corner of the box. The precipitation shadow caused by Mount Discovery and adjacent topographic features is the dark blue patch north of Discovery and south of Ross Island. B: Inset with enlargement of the Dry Valleys region (location of inset is indicated by black box in left panel, A). This figure is available in colour online at www.interscience.wiley.com/ijoc

near the coast, Explorer's Cove, in 2004, but most values were 10–50 mm weq. Modeling shows increased precipitation with elevation and with proximity to the coast. At the mouth of Taylor Valley, where elevation does not play a role, we infer that increased precipitation is due to enhanced atmospheric ascent. Local uplift is caused by converging air flow at the mouth of Taylor Valley as it decelerates and 'piles up', and perhaps is augmented by an onshore component from near surface wind that blows from the Ross Ice Shelf. The low-level precipitation-bearing clouds that result may migrate up valley owing to an upper-level return flow but are blocked by the Nussbaum Riegel, preventing further movement inland. In Wright and Victoria Valleys, modeling suggests that precipitation minima are coincident with the lowest elevations. At Lake Vanda, our snow accumulation values are 18 and 30 mm weq, almost an order of magnitude higher than that measured in the late 1960s–1970s of 3–4 mm weq (Keys, 1980; Bromley, 1985). As described below, snow accumulation has increased in recent years and our larger values at Lake Vanda probably reflect this increase.

Katabatic winds apparently contribute about half of the accumulated snow, indicating that precipitation does not exceed about 50 mm weq in the valley bottom. We infer that katabatic winds must transport local snow from the nearby valley walls onto neighboring valley bottoms because the ratio of precipitation to accumulation is about the same at all sites, ~50%. If the snow was transported from the Antarctic Plateau, then the wind drift snow should be greater inland, closer to the ice sheet (e.g. Lake Bonney), and diminish toward the coast. Keys (1980) argues for the occasional severe katabatic event

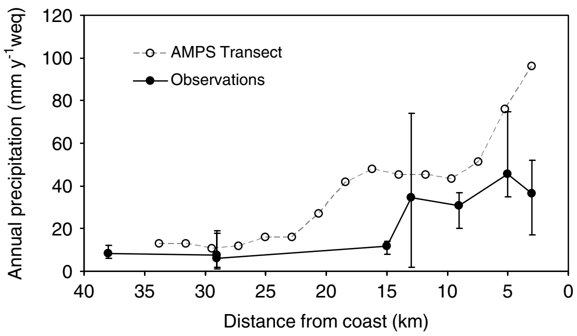

Figure 8. Measured *versus* modeled precipitation for Taylor Valley. The largest precipitation values are closest to the coast and the least are furthest from the coast.

that transports snow from the Antarctic Plateau to the western end of Wright Valley. These events may indeed take place, but do not appear to be a common source of snow to the valleys.

Seasonality of snowfall depends on location in the valleys. No seasonality is apparent at Lake Bonney and a distinct winter maximum appears at Explorers Cove, when katabatic winds are strongest and most frequent (Nylen *et al*., 2004). Year-round observations for 3 years at Lake Vanda showed more frequent snowfall events in summer but the monthly magnitude was no different from winter months (Keys, 1980; Bromley, 1985). Our mass balance measurements on the glaciers (made twice a year) show that either winter or summer can accumulate the most snow. Because snow accumulation and precipitation are low, any one snow storm can shift maximum accumulation between seasons as reflected in the seasonal results on the glaciers (Fountain *et al*., 2006).

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Regional moisture and precipitation in this region results from low-pressure systems migrating from the northern latitudes (Bromwich and Wang, 2008; Carrasco *et al*., 2003; Bromley, 1985). We had hypothesized that more snow is received when the edge of the sea ice is closer, providing a local moisture source; however, we did not find any significant correlations. The sea ice edge is closest to the valley in early March and rapidly advances northward in April (Kimura, 2007), yet maximum precipitation at Explorers Cove occurs in June.

Annual snow accumulation decreases inland from the coast; the accumulation gradient is more steep, -2.7 mm km^{-1}, near the coast and less steep inland, -0.3 mm km^{-1}. The change in gradient is associated with a large 800-m-tall hill, the Nussbaum Riegel, which occupies the center of the valley. The abrupt decrease in snow accumulation inland (west) of the Nussbaum Riegel has been frequently observed in the field and is reflected in the abrupt rise of the glacier equilibrium line altitude (where snow accumulation equals ablation resulting in no mass change) (Fountain *et al*., 1999a). The change in precipitation gradient probably reflects the enhanced snowfall at the valley mouth, described previously, and the barrier formed by the Nussbaum Riegel that blocks the low clouds from migrating up valley. We commonly observe clouds on the coastal (eastern) side of the Riegel, whereas the inland (western) side is cloud free. The lower accumulation gradient beyond the Nussbaum Riegel (-0.3 mm km^{-1}) is very close to that observed in Wright Valley (-0.4 mm km^{-1}) and may represent accumulation from those few large snow storms that override the barrier imposed by the TransAntarctic Mountains and penetrate the deeper portions of the valleys, as well as snow blown to the valley bottom by the katabatic winds.

Modeling shows that in the valley bottom sublimation (not shown) far exceeds precipitation, consistent with the limited duration of snow cover shown in Figure 6. The persistence of winter snow is greatest at Lake Vida in Victoria Valley, nearly all winter, and least at Lake Bonney in Taylor Valley, a few days to a week. This difference is probably due to greater snowfall at Lake Vida and due to the large local inversions that protect the snow from erosion and sublimation by winter katabatic winds. Nylen *et al*. (2004) showed that Lake Bonney had the most frequent katabatic events and Lake Vida had fewest, and hypothesized that strong inversions over Lake Vida resisted erosion by all but the strongest events. Most likely, the combination of snow depth and katabatic winds controls the persistence of snow-covered ground elsewhere in the valleys. How much of the moisture from the snow enters the ground is unclear because much of it sublimates prior to melting (Chinn, 1981). Indeed, snow patches that persist well into summer (December) largely sublimate rather than melt (Gooseff *et al*., 2003). Therefore, these values of precipitation can only be used as a maximum upper bound on moisture recharge to the soils.

7. Conclusions

The MDV is in a precipitation shadow of the TransAntarctic Mountains, blocking precipitation bearing cyclonic storms that migrate southward over the Ross Sea. Our measurements show that the MDV is, indeed, a polar desert, with annual precipitation not exceeding 50 mm weq. However, a precipitation maximum is limited to the coastal margin of Taylor Valley and is caused by the winter convergence of local wind regimes that produce localized uplift. The minimum recorded value of snow accumulation is 3 mm weq at Lake Bonney of which we estimate two-thirds is precipitation and the remaining one-third is snow drift. Snow accumulation decreases inland and a particularly strong gradient exists in Taylor Valley, probably controlled by local meteorological conditions that produce uplift and increased snowfall at the coastal end of the valley.

Wind drift snow is an important contribution to the total snow accumulation on the valley floors. Katabatic winds transport snow from the valley walls and glaciers, and occasionally from the Antarctic Plateau, that roughly doubles the snow accumulation on the valley floor. The persistence of snow on the valley floors is greater in winter than in summer, as one might expect, but is subject to large spatial variability. Winter snow lasts the longest at Lake Vida where it persists nearly all winter, but it rarely lasts more than a week at Lake Bonney. The difference is probably due to the mass of snow accumulated at each site and the frequency of katabatic events that erode and sublimate the snow cover.

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