Performance of Hot Plate for Measuring Solid Precipitation in Complex Terrain during the 2010 Vancouver Winter Olympics

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ABSTRACT

Solid precipitation intensity, snow density, wind speed, and temperature were collected from November 2009 to February 2010 at a naturally sheltered station located at an altitude of 1640 m MSL on Whistler Mountain in British Colombia, Canada. The snowfall was measured using the instruments OTT Pluvio; the Yankee Environmental Systems, Inc., hot plate (HP); and the Vaisala FD12P (optical weather sensor). The snow amount and density were also measured manually daily. The observed wind speeds were in the range 0–4.5 m s\(^{-1}\) with a mean value of 0.5 m s\(^{-1}\). Based on this study, the HP overestimated the snow amount by about a factor of 2 as compared to the Pluvio measurements. Further data analysis using the raw output HP data suggests that this was because of false precipitations produced, particularly by the downslope flows in the complex terrain when the wind speeds were relatively stronger. This false precipitation varied from 0.9 to 1.3 mm h\(^{-1}\) with two peaks at 0.1 and 0.3 mm h\(^{-1}\) depending on wind speed—the larger peak being at higher wind speeds. Since the observed wind speeds were relatively calm, setting the correction factor to 0.15 mm h\(^{-1}\) gave reasonable values as compared to the Pluvio data. The difference between the corrected HP and Pluvio accumulation data varied from 16% to 3% depending on wind speed. The observed snow density in January 2010 varied from 0.04 to 0.32 g cm\(^{-3}\) with a mean value of 0.08 g cm\(^{-3}\). The snow amount measured using the corrected HP data agreed well with the manually measured values with a correlation coefficient of 0.93.

1. Introduction

Precipitation plays an essential role in our planet by controlling the hydrological cycle and hence affecting climate and weather. Thus, accurate measurement of precipitation is necessary for understanding its spatial and temporal variability. Such measurement also has a critical importance for development and validation of the numerical weather prediction, climate models, and satellite and radar remote sensing algorithms. It is now well recognized that accurate determination of solid precipitation is particularly challenging. This is because the traditionally used snow gauges, such as the OTT Pluvio and Geonor, can be affected by many factors, including wind- and evaporation-induced loses (Rasmussen et al. 2012) and wet snow capping (Boudala et al. 2014; Rasmussen et al. 2012). Although currently available automatic gauges, such as the Geonor and Pluvio, can provide short-term precipitation intensity, they are not sensitive enough to provide reliable intensities at short temporal scales (in minutes) (Boudala...
There are other kinds of instruments that can measure precipitation, such as the hot plates (Rasmussen et al. 2011) and forward-scattering probes, such as the FD12P (Haavasoja et al. 1994). These instruments are much more sensitive than the traditional gauges, but these gauges are relatively new and untested. In particular, these instruments are not well tested in a complex terrain environment, where it is particularly challenging to measure precipitation.

The hot plates (HP) are increasingly used for measuring precipitation (Rasmussen et al. 2011, 2012). The HP is attractive for measuring precipitation because of its portability and associated straightforward basic principles employed for determination of precipitation. However, the HP can be affected by wind and other factors, such as radiation, and hence requires some corrections (Rasmussen et al. 2011). This study will explore the validity of the internal algorithms used in the Yankee Environmental Systems, Inc., hot plates for determination of solid precipitation in a complex terrain environment by comparing it with a much more traditional gauge—the OTT Pluvio—and the FD12P in a naturally sheltered area to minimize the effect of wind.

2. Observations

The Pluvio and FD12P data presented here are also discussed in Boudala et al. (2014). Here, a brief description of the dataset will be given. Precipitation intensity and other standard meteorological data were collected at a station (called VOA) located at 1640-m height MSL on Whistler Mountain (see Fig. 1), British Colombia, Canada. The data were collected as part of the Science and Nowcasting Olympic Weather for Vancouver 2010 (SNOW-V10) project. More details about this site are also given in Isaac et al. (2014) and Boudala et al. (2014). Since the location is naturally sheltered by trees, it is an ideal place to measure precipitation (Boudala et al. 2014; Rasmussen et al. 2012). The VOA station had three platforms. One of them was part of the Olympic observation network site that was set up by the Pacific and Yukon Region (PYR) of Environment Canada, where the OTT Pluvio gauge was placed. The second, which was placed a few meters away from the PYR platform, was a SNOW-V10 research platform, where the FD12P, hot plate, and other optical instruments were installed. The third platform, where the manual measurements were made, was located approximately 10 m away from the other two platforms. This is the Whistler Mountain long-term station. One-minute-averaged data from the Pluvio, hot plate, FD12P, and the other instruments were collected in November 2009 and from January to February 2010. Manual snow depth and occasional snow density measurements were also performed daily. For the comparisons between the Pluvio and the other instruments such as the HP and FD12P, only the periods when 1-min-averaged Pluvio data were available were included in this study, since there have been some dates that the data were unavailable due to power loss and some communication problems (see Boudala et al. 2014). There were no 1-min-averaged Pluvio and HP data for the entire month of December 2009, and hence it is not included in this study.

Instrumentations

Detailed descriptions of instruments used in this study are given in Boudala et al. (2014); here, only brief descriptions will be given. The precipitation data were measured using the OTT Pluvio 1 gauge, the Vaisala FD12P, and the Yankee hot plate.

The Pluvio 1 is a weighting gauge. It is equipped with a container with a collecting area of 200 cm$^2$. The weight of the precipitation collected in the container is measured with an electronic weighing cell at a resolution of 0.1 mm with a capacity of measuring up to 1000 mm. The OTT Pluvio calculates the precipitation weight every 6 s using a special intrinsic filter algorithm to correct for wind effects. The measured precipitation intensity or amount is reported every minute. The Pluvio gauge was initially equipped with a collecting
ring-heating device to avoid the snow capping problem and a Tretyakov Shield supplied by the OTT to minimize the effect of wind. The shield created some snow capping problems by accumulating the heavy, wet snow, and hence the shield was removed to minimize these problems. However, the capping persisted because of the fact that the heated ring worked poorly as a result of heavy and wet snow. To avoid these problems, the gauge was wrapped around with a thermostatically controlled water pipe heat tape that was automatically controlled to keep the surface close to \(2^\circ C\) when the outside temperature was less than \(2^\circ C\) (see Boudala et al. 2014). Boudala et al. (2014) have shown that the snow amount measured with the Pluvio at this naturally sheltered location agreed quite well with manual snow measurements with an overall average difference of only 4%, indicating that the Pluvio heating technique mentioned above worked well and that the effect of wind was relatively minor, which gave overall confidence toward testing the hot plates against the Pluvio measurements.

The Vaisala FD12P is an optical instrument (Haavasoja et al. 1994). It measures precipitation intensity, precipitation type, and visibility. The probe combines forward scattering and an analog capacitive surface that senses precipitation falling on it to determine visibility and the liquid water equivalent precipitation rate.

This study mainly focuses on the hot plates, and hence detailed descriptions of this instrument are given in section 3.

### 3. Description of the hot plate precipitation gauge

Figure 2 shows an image of the HP that is similar to the one shown in Rasmussen et al. (2011). As shown in the figure, two identical upper and lower circular plates that are made of aluminum are sandwiched together, while the upper collector plate is facing upward (Fig. 2a), and the other facing downward, which is referred to as the reference plate (shown in Fig. 2b for clarity). The upper plate is exposed to falling precipitation particles, but the lower plate is facing downward so that it is protected from collecting the falling hydrometeors. The diameter of these plates is 13 cm. The smaller two internal concentric circles are designed to prevent snow particles from sliding off the upper plate. The two plates are heated to the same temperature \((T_p > 75^\circ C)\), but the insulation between the two plates does not allow heat transfer. The cooling rate of the upper plate as compared to the lower plate during precipitation can be related to the melted equivalent precipitation intensity. Detailed information on the measurement principles of this instrument is given in the following sections.

#### a. Measurement principles

The hot plates measure precipitation following the basic principles of heat transfer.

Heat loss from the upper plate due to both forced convection and precipitation in joules per second \((J s^{-1})\) is given as

\[
P = \alpha(T_p - T_a) + L A_p \rho_w R,
\]

and the heat loss from the lower (reference) plate is due to only forced convection and it is given as

\[
P_r = \alpha'(T_r - T_a),
\]

where \(R\) is the precipitation intensity in millimeters per hour; \(L\) is the energy necessary to heating and evaporation of the water; \(A_p\) is the area of the plate; \(\rho_w\) is the density of water; \(T_p\) and \(T_r\) are the temperature of the upper and lower reference plates, respectively; \(\alpha\) and \(\alpha'\) are the bulk convective heat transfer coefficients for the upper and lower reference plates, respectively; and \(T_a\) is the temperature of the ambient air. The power difference between the plates \((\Delta P = P - P_r)\) can be obtained by subtracting Eq. (2) from Eq. (1) as
\[ \Delta P = \alpha(T_p - T_a) - \alpha'(T_r - T_a) + LA_p \rho_w R. \]  

(3)

As indicated in Eq. (3), there are two forced convective heat loss terms associated with upper and lower plates, and the last term is associated with precipitation. Note that here the effects of sensible and radiative cooling–heating are neglected. During precipitation, it is generally assumed that Eq. (3) can be approximated as

\[ \Delta P \approx LA_p \rho_w R, \]  

(4)

since the two convection terms cancel each other out. When there is no precipitation \((R = 0)\), it is also expected that

\[ \Delta P = \alpha(T_p - T_a) - \alpha'(T_r - T_a) \approx 0. \]  

(5)

Provided this assumption holds, the melted equivalent precipitation rate in millimeters per hour for rain \((R_r)\) and snow \((R_s)\) can be calculated as

\[ R_r = \frac{(1000 \text{ mm m}^{-3})(3.6 \times 10^3 \text{ s}) \text{ h}^{-1}}{L_v A_p \rho_w} \Delta P = f_L \Delta P, \]  

(6)

\[ R_s = \frac{(1000 \text{ mm m}^{-3})(3.6 \times 10^3 \text{ s}) \text{ h}^{-1}}{L_s A_p \rho_w} \Delta P = f_S \Delta P, \]  

(7)

where \(L_v\) and \(L_s\) are the latent heat of vaporization and sublimation, respectively. The average values of these parameters may be taken as \(L_v = 2.58 \times 10^6 \text{ J kg}^{-1}\) and \(L_s = 1.13L_v\) (Korolev et al. 1998). The area of the hot plate used in this study is \(13.267 \times 10^{-3} \text{ m}^2\) and the last term is associated with precipitation. Note that here the effects of sensible and radiative cooling–heating are neglected. During precipitation, it is generally assumed that Eq. (3) can be approximated as

\[ \Delta P \approx LA_p \rho_w R, \]  

(4)

and hence \(f_L\) is used. For temperatures between 0°C and 4°C, a linear combination of \(f_L\) and \(f_S\) was used.

In reality, however, the assumption given in Eq. (5) does not always hold due to a heating imbalance between the two hot plates that produces false precipitation, and hence some correction is required. Based on Rasmussen et al. (2011), this correction factor has a wind speed \((w)\) dependence. Therefore, the expression for deriving precipitation using the hot plates can be given as

\[ R = f \Delta P - R_{\text{cor}}, \]  

(8)

where \(R_{\text{cor}}\) is the correction factor in millimeters per hour. This can be determined using Eq. (6) or Eq. (7) depending on temperature, knowing that no precipitation intensity has been detected, but the power difference \(\Delta P\) is nonzero. With this correction factor, it is expected that when precipitation is not detected, the right-side terms cancel each other and hence \(R = 0\). This will be discussed in more detail in section 5.

The expression given in Eq. (8) also needs to be corrected for collection efficiency \((\text{CollEff})\) of the upper plate for solid precipitation. Based on Rasmussen et al. (2011), this parameter depends on wind speed and it is given as

\[ \text{CollEff} = 1 - 0.08 w_{s10} \quad w_{s10} < 10 \text{ m s}^{-1} \]

\[ = 0.2 \quad w_{s10} > 10 \text{ m s}^{-1}, \]  

(9)

where the wind speed is measured at a 10-m height \((w_{s10})\). Hence, after correction for collection efficiency the precipitation intensity is given as

\[ R' = R/\text{CollEff}. \]  

(10)

This collection efficiency factor was developed by Rasmussen et al. based on data collected using the hot plate and a Geonor gauge placed in a DFIR at an open and windy location. Therefore, it is generally expected to work better in an open field, where the airflow impacting the hot plates is nearly horizontal. Furthermore, since the wind speed is normally derived based on \(P_r\) and \(T_n\), which are generally measured at 2-m height, some adjustment for wind speed is also required if Eq. (9) is to be used. It is expected that Yankee may have a similar collection efficiency algorithm that uses wind speed measured at a 2-m height. This equation was not available, and Eqs. (9) and (10) were not tested in this paper. However, CollEff is available as an output variable from the probe.

b. Wind speed and collection efficiency

Figure 3a shows CollEff and gauge-level wind speed \((w_{sHP})\) reported by the HP, and Fig. 3b shows \(w_{sHP}\)
plotted against 2-m observed wind speed at the VOA site. As indicated in the figure, the HP probe overestimates the wind by more than a factor of 2 under some conditions, indicating that the HP wind algorithm in complex terrain and sheltered locations does not work well. Therefore, the collection efficiency derived using \( w_{\text{shp}} \) would overestimate the precipitation intensity, particularly at higher wind speeds where the collection efficiencies are smaller than 0.8. As indicated in Fig. 3b, the 2-m observed wind speeds were generally very low (<4 m s\(^{-1}\) with a mean value of 0.5 m s\(^{-1}\)), and hence the associated collection efficiencies would be much higher than the values shown in Fig. 3a.

c. Correction factor

As discussed earlier, when winds preferentially cool the top plate more than the bottom plate, false indications of precipitation can occur. This may happen when downslope flow occurs at the site during the evening hours, for instance, or during down-valley flow over the mountain from the north. To identify the meteorological conditions influencing the HP to produce false precipitation events, false precipitation intensity calculated using Eqs. (6) and (7) is plotted in Fig. 4 against 2-m observed wind speed (Fig. 4a), 2-m observed wind speed and direction (Fig. 4c), and temperature (Fig. 4d). The frequency distribution of the false precipitation for several 2-m observed wind speed intervals is also given in Fig. 4b. For the data in this plot, the entire datasets mentioned in section 2 were used, but only when the FD12P detected no precipitation. There are some indications that the false precipitations that are associated with wind directions between 100° and 200°, and to the lesser extent from 300° to 350° are significantly higher when the wind speeds were relatively stronger (Fig. 4b). These wind directions are consistent with downslope flow from the mountain (see Fig. 1). Hence, this suggests that both wind direction and wind speed are important factors, but as indicated in Fig. 4a, there is no significant correlation of false precipitation with wind speed alone. Based on the frequency distribution of false precipitation, there is also some indication that for larger wind speeds (\( w_z > 2 \text{ m s}^{-1} \)), the false precipitations are larger. The false precipitation varies from −0.9 to 1.3 mm h\(^{-1}\) with peaks at 0.1 mm h\(^{-1}\) and 0.3 mm h\(^{-1}\) depending on wind speed (Fig. 4b). Although the false precipitation showed some systematic dependence on wind speed and wind direction, the dependence is not easily parameterized as a function of any of the meteorological parameters discussed above. Hence, in this study, setting the correction factor \( R_c \) to be a constant value of 0.15 mm h\(^{-1}\) gave reasonable results, and this is discussed in the following section.
4. Comparisons of precipitation

a. HP against Pluvio and FD12P

Figure 5 shows comparisons of solid precipitation (SP) amount measured with the Pluvio gauge, FD12P, and reported directly by the HP (HPrep), calculated using the raw HP data and Eq. (7) without any correction for collection efficiency or false precipitation (HPcal) and with correction for false precipitation (HPcor) using Eq. (8) without the collection efficiency correction, assuming the correction factor is 0.15 mm h\(^{-1}\) (Fig. 5a). The corresponding temperature and wind speed are shown in Fig. 5c. Similar plots to Figs. 5a and 5c are shown in Figs. 5b and 5d, respectively, but for 2-m observed wind speed less than 1 m s\(^{-1}\). The data were collected in November 2009 and from 1 January to 17 February 2010. In Fig. 5, one data point represents 1 min of averaged data. The total SP amounts based on the HP were 343, 259, and 175 mm for HPrep, HPcal, and HPcor, respectively. For the same period, the SP measured using the Pluvio and FD12P were 150.4 and 111 mm, respectively. The ratios HPrep/Pluvio, HPcal/Pluvio, HPcor/Pluvio, and FD12P/Pluvio were 2.3, 1.72, 1.16, and 0.74, respectively. Hence, the total SP amount reported by the HP was larger than the value measured with Pluvio by a little more than a factor of 2. However, after correction for false precipitation (HPcal − 0.15), the calculated precipitation is quite close to the value measured with Pluvio with only a 16% difference. The FD12P seems to underestimate the SP by about 26% as compared to the value measured by Pluvio. Note that the collection efficiency effect on HP was not included in the calculation of HPcal, but it can be seen by comparing the reported values that include this effect, and the calculated values before correction and after correction (Fig. 5a). The effect of wind is relatively small as compared to the effect of false precipitation. Based on the results found here, the effect of wind via collection efficiency was close to 10%. It can be seen in Fig. 5a, however, that the effect of false precipitation on HP increases with increasing time, and hence for shorter time periods, the effect of collection efficiency may dominate if the wind is sufficiently large.
When the data were segregated based on the 2-m observed wind speed (<1 m s\(^{-1}\)) as in Figs. 5c and 5d, the reported and calculated SP amounts were 263, 198, and 132.6 mm for \(H_{\text{Prep}}\), \(H_{\text{Pcal}}\), and \(H_{\text{Pcor}}\), respectively, and the SP amounts measured by Pluvio and FD12P were 128 and 86 mm, respectively. The ratios \(H_{\text{Prep}}/\text{Pluvio}\), \(H_{\text{Pcal}}/\text{Pluvio}\), \(H_{\text{Pcor}}/\text{Pluvio}\), and \(\text{FD12P}/\text{Pluvio}\) were for 2.1, 1.55, 1.03, and 0.7, respectively. The ratios changed by approximately 10%, indicating the effects of wind speed were relatively weak, but the \(H_{\text{Pcor}}\) value was much closer to the value measured by the Pluvio gauge with only a 3% difference as compared to a 16% difference when all the wind speed values were included in the analysis.

These results suggest that the internal algorithms used in HP may not be applicable for a sheltered, complex terrain environment due to several factors. These include the overestimation of wind speed by the HP when it is sheltered, which affects the collection efficiency algorithm, and downslope flow, which may induce false precipitation. Nonetheless, a simple correction factor for false precipitation and neglecting the effect of collection efficiency seems to remove most of the bias.

b. HP against manual measurements: Snow depth and snow density

As mentioned earlier in the introduction, there were manual measurements of snow depth and snow density at the VOA site. The manual snow depth readings taken daily at VOA were done by a permanent tall ruler left in the snow, and the data are available for all winter seasons. The snow density measurements were made sporadically once per day during the Olympics period in December 2009 and January 2010 using a Snowmetrics T1 tube sampler and a hanging weighing scale. The Snowmetrics T1 tube sampler is about 30.5 cm in length and about 5.5 cm in diameter, and marked with a ruler in metric and English graduations. The snow depth is measured by inserting the tube into freshly fallen snow typically collected on a snowboard. The amount of snow in the tube was scooped with a help of a scrapper and secured by closing the tube with a lid.
The weight of the snow as water equivalent (SWE) in millimeters was determined by weighing the tube and the snow using the hanging scale.

In this study, we have included snow depth measurements taken in November 2009 and January–February 2010. However, as stated earlier, there were no 1-min-averaged HP data for the entire month of December 2009, and hence the snow density information cannot be used. For the rest of the period, snow density measurements were only made in January, and for November and February the mean snow density of 12:1 ratio measured in January was assumed (see the discussion below) for converting the snow depth to SWE. Using measurements of snow depth \( D_s \) in centimeters and associated SWE in millimeters, the snow to SWE ratio (SWR) can be computed as

\[
\text{SWR} = \frac{10D_s}{\text{SWE}},
\]

(11a)

and the bulk snow density is given as

\[
\rho_s = \frac{\rho_w}{\text{SWR}}.
\]

(11b)

where \( \rho_w \) is the density of pure water, which is 1.

Figure 6 shows the daily variation of observed SWR (Fig. 6a) and associated snow density (Fig. 6b), and temperature (maximum, minimum, mean) (Fig. 6c) for January 2010. The values of SWR range from 3.1 to 26 (Fig. 6a) with an associated range of densities of 0.32–0.04 g cm\(^{-3}\) (Fig. 6b). The calculated mean values for SWR and \( \rho_s \) were 12.3 and 0.08 g cm\(^{-3}\), respectively. Traditionally, the average SWR is assumed to be 10 for most meteorological applications based on a Canadian study reported by Potter (1965) (see Roebber et al. 2003 for discussions), and this would imply a snow density of 0.1 g cm\(^{-3}\), which is very close to the mean \( \rho_s \) value found in this study, but slightly higher. As would be expected, the two higher snow density days (the 10th and 11th) are associated with relatively warmer temperatures with a mean value near 0°C (Fig. 6c).

Figure 7 shows the daily accumulation rate of the SWE measured manually and using the hot plate (HP\(_{\text{cor}}\)) (Fig. 7a), the total accumulation starting on 14 November 2009 to the end of November and from January to February 2010 (Fig. 7b), the observed daily averaged minimum and maximum temperatures (Fig. 7c), and the HP\(_{\text{cor}}\) data plotted against the manual measurements (Fig. 7d). The manual snow depth measurements were converted to SWE with the help of the snow density for...
the January data as discussed above, and when this information was missing, the mean snow density of 0.08 g cm$^{-3}$ was used for the rest of the data. The inclusion of the snow density information for January data improved the agreement with the corrected HP data by increasing the correlation coefficient from 0.68 to 0.84 and decreasing the mean bias from 0.34 to 0.22 mm. A total of about 76 daily manual measurements are included in Fig. 7. The daily averaged temperature varied from a maximum of 4.5°C in February to −12°C. The total amounts of solid precipitation measured using the HP and the manual method were very close—697 and 721 mm, respectively. The two datasets correlated well with a correlation coefficient ($r$) of 0.93 with a mean difference (MD) of 0.32 (mm), but the data also shows occasional discrepancy at the lower end of the data (Fig. 7d) that may be associated with uncertainties in snow density and manual snow depth measurements and other factors, such as uncertainty in the HP precipitation algorithms.

5. Summary and conclusions

Solid precipitation intensity, snow depth, snow density, and relevant meteorological parameters were collected at a station that is in a naturally sheltered area located at an altitude of about 1640 m MSL on Whistler Mountain in British Colombia, Canada. The precipitation intensity was measured using the Pluvio, a hot plate, and FD12P instruments. The data analyzed in this paper were collected from November 2009 to February 2010, which is the same dataset reported in Boudala et al. (2014). The wind speeds measured during this period were in the range 0–4.5 m s$^{-1}$ with a mean value of 0.5 m s$^{-1}$. This study was particularly intended to explore the validity of the internal algorithms used in the Yankee hot plates for determination of solid precipitation in a complex terrain environment by comparing it with much more traditional gauges: the OTT Pluvio and the FD12P. Based on this study, there is an indication that the hot plate overestimates solid...
precipitation amount by approximately a factor of 2 as compared to the Pluvio gauge measurements. There are some indications that this is because of false precipitations produced by the downslope flow in the complex terrain, particularly when the wind speeds were stronger. This false precipitation varied from −0.9 to 1.3 mm h\(^{-1}\) with two peaks at 0.2 and 0.4 mm h\(^{-1}\) depending on wind speed—the larger peak being at higher wind speeds (\(w_s > 2 \text{ m s}^{-1}\)). Nonetheless, since the observed wind speeds were relatively calm, neglecting the collection efficiency and setting the correction factor to 0.15 mm h\(^{-1}\) gave reasonable values as compared to the Pluvio, particularly at lower wind speeds. After the correction was applied, the difference between the hot-plate and the Pluvio accumulation varied from 16% to 3%, the 3% difference was being at the lower wind speeds (\(w_s < 1 \text{ m s}^{-1}\)).

The snow density measured in January 2010 varied from 0.04 to 0.32 g cm\(^{-3}\) with a mean value of 0.08 g cm\(^{-3}\). The hot plate data were also compared to the daily manual measurements, and the agreement was good with a correlation coefficient of 0.92 with a mean difference just 0.32 mm.

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