RESEARCH ARTICLE

DIAL and GNSS observations of the diurnal water-vapour cycle above Iqaluit, Nunavut

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Abstract
Atmospheric water vapour is the dominant gas in the greenhouse effect and its diurnal cycle is an essential component of the hydrological cycle. High-quality water-vapour measurements are critical observations for numerical weather prediction (NWP) models, which encounter difficulties in the Arctic due to its unique weather. Accurate measurements of the diurnal water-vapour cycle can improve precipitation rate and type predictions. In this study, we use a pre-production Vaisala broad-band differential absorption lidar (DIAL), installed in Iqaluit, Nunavut (63.75°N, 68.55°W), to calculate seasonal height-resolved diurnal water-vapour cycles from 100–1,500 m altitude. We also calculate the surface water-vapour mixing ratio and integrated water-vapour (IWV) diurnal cycles using co-located surface station and global navigation satellite system (GNSS) measurements. We find that the first 250 m of the DIAL water-vapour mixing ratio height-resolved diurnal cycle agrees within 0.02 g kg⁻¹ with the surface-station amplitudes, and within 0–2 hr in phase. DIAL diurnal-cycle \( R^2 \) values are close to 1 at the surface, and between 0.25 and 0.95 depending on the altitude and season. The phases of the diurnal cycle shift and the amplitudes increase with altitude. In the summer, all instruments observe a strong 24-hr cycle. The amplitude of the 24-hr component decreases with the solar cycle, such that the 12-hr component begins to influence the total cycle significantly by the winter. The IWV has a large 12-hr component throughout the year, which could be due to the superposition of two diurnal components at different altitudes or increasing contributions at altitudes higher than those measured by the DIAL. We have shown that using DIAL measurements to observe height-resolved diurnal cycles provides a deeper understanding of the diurnal cycle than that achieved from GNSS and surface measurements alone.

KEYWORDS
Arctic, boundary layer, DIAL, diurnal, GNSS, lidar, water vapour

1 INTRODUCTION

Water vapour is one of Earth’s most important trace constituents. The diurnal water-vapour cycle, in particular, is a crucial component of the larger hydrological cycle. The diurnal cycle is an atmospheric tide and has two main components: the 24-hr diurnal and 12-hr semidiurnal oscillations. The diurnal component is mostly driven...
by evapotranspiration via the absorption of solar radiation at the surface, but also by atmospheric large-scale vertical motion, atmospheric moisture convergence and precipitation, and vertical mixing in the boundary layer (Dai et al., 1999; 2002; Morland et al., 2009). Previous studies have also discussed these drivers in relation to the total diurnal cycle, but no particular driver has been attributed to the semidiurnal cycle specifically. While the diurnal-cycle drivers are well known, the drivers of the semidiurnal cycle are largely undiscussed in the literature.

Given the diurnal cycle’s dependence on multiple drivers, measurements of the diurnal water-vapour cycle are an extremely useful tool for numerical weather prediction (NWP) model validation and verification. Accurate modelling of atmospheric water vapour is required to support their radiative transfer and convection algorithms and provide correct forecasts (Dai et al., 1999; Fabry, 2010; Fabry and Sun, 2010; Bannister et al., 2020). Improvements in NWP modelling of the diurnal cycle will lead to subsequent improvements in precipitation and cloud-type predictions (Dai et al., 2002; Trenberth et al., 2003; Dai and Trenberth, 2004; Hocke et al., 2017).

Until the beginning of the millennium, obtaining estimates of the atmospheric diurnal water-vapour cycle was difficult due to the lack of instrumentation with sufficient temporal resolution. While radiosondes were available, most were (and still are) launched only twice per day, and have not been timed to occur with the maximum and minimum diurnal water-vapour anomalies. Dai et al. (2002) demonstrated that observations from radiosondes launched twice per day can result in uncertainties in the mean diurnal water-vapour cycle, as the standard launch times of 0000 and 1200 UTC may not capture the full amplitude of the cycle. In the 1990s, ground-based global navigation satellite systems (GNSS) became an established method of measuring total column or integrated water vapour (IWV: Bevis et al., 1992; 1994). Due to its high temporal resolution (hourly or subhourly), GNSS has become a preferred method of calculating diurnal IWV cycles (Bouma and Stoew, 2001; Dai et al., 2002; Wang and Zhang, 2008; Galisteo et al., 2011; Hocke et al., 2017).

However, ground-based GNSS measurements cannot provide insight into the vertical structure of the diurnal cycle. Very few studies have been conducted regarding height-resolved diurnal water-vapour cycles. Dai et al. (2002) used three-hourly radiosonde measurements at the Atmospheric Radiation Measurement (ARM) programme site in Lamont, Oklahoma to measure vertical diurnal structures; however, very few sites launch radiosondes with such frequency. Microwave radiometers have been used to measure height-resolved diurnal cycles, particularly in the mesosphere (Haefele et al., 2008; Hallgren and Hartogh, 2012; Scheiben et al., 2013). Louf et al. (2015) and Wang et al. (2002) demonstrated that microwave radiometers can be used to calculate tropospheric diurnal water-vapour cycles as well. Although newer models are being developed with higher vertical resolution, most microwave radiometers have relatively low vertical resolution in the troposphere, of the order of several hundred metres, which is too low to resolve much vertical structure in the diurnal cycle.

Laser radars, or lidars, are another option for height-resolved diurnal water-vapour observations. Lidars typically have two methods of measuring water vapour. The first is through Raman scattering, otherwise called Raman lidar. The second is via differential transmission, or the Differential Absorption Lidar (DIAL) technique. Both techniques have their advantages and disadvantages, but for the purposes of this study we will focus on the DIAL technique. The DIAL technique has been applied to water-vapour measurements since the early 1990s for both ground- and air-based lidars (South et al., 1998; Wulfmeyer and Bösenberg, 1998; Wagner and Prüß, 2002; Spuler et al., 2015). Water-vapour DIALs are now starting to be developed for automated and operational use (Spuler et al., 2015; Newsom et al., 2020). Operational water-vapour DIALs have the advantage of running continuously and can run unattended, alone or as part of a larger network, in a variety of climates and weather regimes. They have excellent temporal resolution of the order of minutes and have vertical resolutions of around 75–100 m. Both Weckwerth et al. (2016) and Newsom et al. (2020) demonstrated that DIALs could resolve diurnal variations and Wulfmeyer et al. (2015) used DIAL water-vapour measurements to resolve turbulence in the boundary layer. DIALs, therefore, make excellent candidates for studying diurnal water-vapour cycles. A more detailed description of the DIAL technique and theory will be discussed in Section 2.3.

Previous studies analyzing the diurnal water-vapour cycle have focused on midlatitude and tropical environments (e.g., Dai et al., 2002; Hanesiak et al., 2010; Galisteo et al., 2011; Louf et al., 2015; Torri et al., 2019). However, the amplitude and phase of the diurnal and semidiurnal cycle can change significantly with respect to latitude and geography. Dai et al. (2002) found that the 24-hr component dominated the total diurnal cycle over most of the contiguous United States; however, the 12-hr component was larger than the 24-hr component at their Alaskan sites. Galisteo et al. (2011) found considerable variability in the diurnal IWV cycle around Spain, due to Spain’s large variation in geography and environments.

The Arctic diurnal water-vapour cycle is poorly understood, as few studies have been conducted there due to the lack of available measurements and the fact that relatively
low water-vapour concentrations make accurate measurements difficult. Some global diurnal IWV studies have included Arctic sites, including Wang et al. (2007) and Wang and Zhang (2008); however, the results are averaged over large latitude bands, such that the amplitudes at individual sites and geographic variability are not discernible. Dai et al. (2002), Bouma and Stoew (2001), and Jakobson et al. (2009; 2014) presented the first GNSS diurnal IWV cycles for individual sites in the United States (including Alaska) and the Baltic Sea region. A surface diurnal water-vapour cycle was also calculated over the Greenland ice sheet by Steen-Larsen et al. (2013) for the summer months during clear skies and calm weather only. Hallgren and Hartogh (2012) used microwave radiometer measurements to calculate mesospheric diurnal water-vapour cycles at the ALOMAR observatory in northern Norway. As yet, no ground-based studies of the diurnal IWV cycle have been published for the Canadian Arctic. Additionally, there appears to be a distinct lack of height-resolved diurnal studies globally. However, height-resolved humidity measurements with high vertical and temporal resolution are extremely important for improving NWP models and have been classified as a priority measurement (National Research Council, 2009; National Academies of Science, 2016).

The Environment and Climate Change Canada (ECCC) Canadian Arctic Weather Science (CAWS) site in Iqaluit, Nunavut (Joe et al., 2020) is an ideal location at which to study the Canadian Arctic diurnal water-vapour cycle. The Iqaluit “super site” is a coastal site located at 63.75°N, 68.55°W at 11 m above sea level. In September 2018, ECCC installed the new preproduction Vaisala DIAL (Mariani et al., 2020; Newsom et al., 2020) at the site to measure water-vapour profiles in the boundary layer. In this study, we use the Vaisala DIAL measurements to calculate height-resolved diurnal water-vapour cycles from 100–1,500 m above ground level. We also compare the DIAL measurements with World Meteorological Organization (WMO) surface station and GNSS measurements that are co-located with the DIAL.

This study presents the first tropospheric height-resolved water vapour and IWV diurnal cycles for the Canadian Arctic, as well as the first diurnal cycles to be calculated with a water-vapour DIAL. While the previously cited studies have focused on the summer months only, we have included all seasons in this study. Section 2 discusses the instruments used in this study. Section 3 describes the methodology behind the diurnal-cycle calculations and harmonic analysis. The surface, DIAL, and GNSS diurnal water-vapour cycles are presented in Section 4. A discussion of the results and a summary and conclusions are then presented in Sections 5 and 6.

2 | INSTRUMENTATION AND DATA PROCESSING

2.1 | Radiosondes

The Iqaluit supersite launches daily radiosondes at 0000 and 1200 UTC, or 1900 and 0700 local time (LT), respectively. We use Vaisala RS92 and GRAW DFM-09 radiosonde measurements in this study to validate the GNSS IWV measurements. The GNSS was installed in 2009, therefore we only use radiosonde measurements from 2009–2020 for comparison. Vaisala RS92 radiosondes were used from January 2009 until November 23, 2018, when ECCC switched to the GRAW DFM-09. We examined the monthly averages and standard deviations of the GRAW and RS92 water-vapour mixing ratio measurements to determine whether a transfer function was required between the two instruments. We found that the average IWV and standard deviations of the measurements did not change, therefore we did not use a transfer function between the two radiosondes when comparing with the GNSS measurements. We compare the two radiosondes and the GNSS measurements in the following section. Because the radiosondes are only flown twice per day, we omitted them from the diurnal-cycle calculations.

2.2 | GNSS

The Iqaluit, Nunavut GNSS unit has been operating almost continuously since 2009. It is managed by National Resources Canada (NRCan) and the International GNSS Service (IGS) and is part of the Canadian Active Control System (CACS). It has seen very little downtime (~2% of the operating time) and the receiver has not been changed since its installation in September 2009. GNSS instruments do not measure IWV directly; instead, they measure the delay time between coded signals passed between the satellites and the ground. The phase of the signal and the delay can be used to calculate the position and velocity of the satellites by referring to a reference signal. The delay in the signal is caused by the density of the atmosphere as well as ionospheric dispersion (Bevis et al., 1992). The atmospheric delay is often divided into two components: the hydrostatic or “dry” atmosphere and the “wet” atmosphere or the delay due to water vapour. Bevis et al. (1992) found that the wet delay could be used to determine the total amount of water vapour in the atmosphere, or IWV, along the path of the GNSS signal. Several concise summaries of how to calculate IWV from GNSS signals can be found in Bevis et al. (1992; 1994), Emardson et al. (1998), Dai et al. (2002), and Jones et al. (2020).
We use updated IGS tropospheric products (including IWV) provided by the University of Nevada Geodetic Lab (NGL: Blewitt, 2019). The NGL has processed measurements from over 17,000 GNSS sites worldwide using data from 1994 to the present. All NGL data are open-access and available through their website (Blewitt, 2019). The tropospheric products are generated using JPL’s GipsyX 1.0 software, JPL’s Repro 3.0 orbits and clocks, and Vienna Mapping Function 1 (VMF1: Kouba, 2007) gridded data and mapping-function parameters. These updated products, as of 2019, provide IWV and its uncertainties directly to the user at 5-min temporal resolution, which is extremely valuable for climate and meteorological studies (Blewitt et al., 2018). Previous IGS GNSS tropospheric products did not include IWV calculations.

Calculating IWV requires an estimate of the mean temperature of the atmosphere, which is typically derived using radiosonde data. Bevis et al. (1994) provided a general relationship between the mean temperature and the surface temperature. The NGL processing calculates mean temperature values through the VMF1 grid every 6 hr, which are then interpolated in space and time to the desired grid points and time. The mean temperatures, along with the constants from Bevis et al. (1994), are used to calculate the final IWV values. While using the constants from Bevis et al. (1994) does introduce some uncertainty into the IWV component, the error in the final IWV value due to the uncertainty in the constants is small compared with the uncertainty in the mean temperature itself and is mitigated by using the VMF1 grid instead of climatological values. Additionally, the errors of the Bevis constants are usually less than 1% of the final uncertainty in the final IWV. The majority of the uncertainty (more than 75%) in the final IWV product is due to the uncertainty in the ZTD derivation (Ning et al., 2016).

To determine whether the Bevis et al. (1994) constants were appropriate for our use, we compared the NGL IWV values with the daily radiosonde IWV values and our own derived IWV values using our own empirical relationship of the same form as in Bevis et al. (1994). The empirical relationship was calculated using the twice-daily radiosondes. We found that both the NGL IWV and our derived IWV product agreed well with the radiosondes with an average bias of −0.42 and −0.46%, respectively. The difference between the two IWV products was negligible, and the NGL IWV bias was slightly lower; therefore, we chose to use the NGL product. There is a bias with respect to the radiosonde present in the winter months (Figure 1). In the winter months, the GNSS measurements are wetter than the radiosonde by roughly 1 mm. This is to be expected, given that radiosondes are known to have dry biases in extremely cold weather due to “time lag” and calibration errors (Miloshevich et al., 2004; 2006; Dirksen et al., 2014).

These cold temperatures are likely the cause of the difference, given that the bias between the two instruments disappears completely as the temperatures increase in the summer. While there is also a slightly smaller wet bias in July and August, it is well within the standard deviation of the measurements.

The IWV measurements taken by the GNSS agree well with other studies and Arctic IWV climatologies. Serreze et al. (1995) used rawinsondes to measure IWV at 70°N and found values between 3 mm in the winter and 16 mm in the summer, which agrees with our observations. Rinke et al. (2009) used satellite measurements to retrieve total water-vapour columns over the Arctic from 1958–2001. They showed that Iqaluit generally had an average IWV between 2 and 3 mm in the winter, and an average of 12–14 mm in the summer (June–August) which is about 1 mm drier than our observations but within the uncertainties of the measurements. Differences in our observations could be due to differences in instrumentation as well as the time periods used.

The NGL tropospheric solutions are processed in daily batches, such that a unique solution for the total zenith delay time series is found every 24 hr. Each 24-hr solution is conducted without any information provided from the previous day’s solution. Unfortunately, this can create a “discontinuity” from day to day at 0000 UTC or 1900 LT if the errors in the GNSS orbit models are large enough.
The average magnitude of the 0000 UTC discontinuity from 2009–2020 for the Iqaluit station is $0.08 \pm 0.62$ mm. Statistically, the average shift value is not significant. There is some seasonality in the size of the discontinuity, with larger discontinuities present in the summer and smaller ones in the winter. However, these shifts are not associated with any particular meteorological process. The subsequent shift in the measurements is not limited to a single point after 0000 UTC, but propagates for 2–3 hr afterwards due to assumed correlation between each subsequent point. Therefore, to smooth the effect of the shift, we first average the GNSS measurements to 1 hr instead of 5 min, which alleviates some of the discontinuities but does not completely remove all of their effects. Then, to mitigate the discontinuity’s effect on the diurnal-cycle calculations further, we developed a simple quality-control filter such that all dates with a discontinuity above a certain threshold or cut-off value are excluded from the analysis (Table 1). Removing the data serves two purposes: we remove dates in which the automatic processing produced large discontinuities and in the process we reduce the average shift for the season closer to zero. It is therefore fair to assume that we are capturing the true average cycle.

The quality-control filter keeps all dates with a shift less than a certain cut-off value. We chose the cut-off value by minimizing the number of days lost and the average shift value, as well as maximizing the goodness of fit of the diurnal-cycle equation. Table 1 shows how many data are removed for each cut-off value in each season, their corresponding average 0000 UTC shift values, and $R^2$ values for each fit. The asterisk indicates which cut-off value was used to produce the final results. In summer, the best fit with the least amount of data loss is for a cut-off value of 0.5. However, almost 40% of the data have to be removed in order to obtain a fit with no visible mean shift and a good $R^2$ value. This is because the summer has higher shift values than the winter and spring. In winter, almost no cut-off is necessary; however, we have applied a cut-off of 0.75 to remove most of the dates with large shifts from the dataset. In spring, we apply a cut-off of 0.5 because, even though the $R^2$ (cut-off = 0.5) is smaller than the $R^2$ (cut-off = 0.25), 30% more data would have been removed and there is no change in the calculated diurnal cycle between the two cut-off values.

### 2.3 | Vaisala DIAL

The Vaisala DIAL preproduction model used in this study is a broad-band water-vapour DIAL. The DIAL was first designed in 2010 and then a prototype was tested in field campaigns in Germany, Finland, and China. Newsom et al. (2020) validated the previous prototype version with radiosonde, Raman lidar, and Atmospheric Emitted Radiance Interferometer (AERI) measurements at the US Department of Energy Southern Great Plains (SGP) site. ECCC acquired the updated preproduction model in 2018 after it underwent testing at the Vaisala headquarters in Vantaa, Finland. The ECCC model was installed at the ECCC Downsview test field for initial comparisons and validation (Mariani et al., 2020) before the final installation at the ECCC Iqaluit, Nunavut supersite (63.75°N, 68.55°W). It has been operating in Iqaluit since September 2018 as part of the CAWS project (Joe et al., 2020). In this study we use DIAL measurements from September 2018–June 2020, during which time the instrument operated with 0% downtime except for the occasional power outage at the site.

Unlike Raman or Rayleigh lidars, the DIAL uses two lasers at two different wavelengths to measure water vapour. The “online” wavelength (911.0 nm) is strongly absorbed by water vapour, and the “offline” wavelength (910.6 nm) is insensitive to water-vapour absorption. One unique difference between the Vaisala DIAL and other DIALs is that each laser’s spectral width is relatively large.

| Cut-off (mm) | Percent removed (%) | Avg. shift (mm) | $R^2$ |
|-------------|---------------------|-----------------|-------|
| Summer: April–August | | | |
| 0.75 | 23.7 | 0.07 | 0.89 |
| $^*0.50$ | 39.1 | 0.05 | 0.90 |
| 0.25 | 64 | 0.02 | 0.76 |
| Fall: September–October | | | |
| 0.75 | 15.1 | 0.06 | 0.89 |
| $^*0.50$ | 30.4 | 0.028 | 0.83 |
| 0.25 | 60.3 | 0.008 | 0.83 |
| Winter: November–January | | | |
| $^*0.75$ | 5.2 | 0.0 | 0.92 |
| 0.50 | 16.2 | 0.0 | 0.89 |
| 0.25 | 42.5 | 0.0 | 0.89 |
| Spring: February and March | | | |
| 0.75 | 3.6 | 0.024 | 0.90 |
| $^*0.5$ | 10.5 | 0.018 | 0.87 |
| 0.25 | 43.3 | 0.001 | 0.89 |

Note: Column 1 is the cut-off value, column 2 is the percentage of dates removed by the filter, column 3 is the average 0000 UTC shift value for the season after the cut-off is applied, and column 4 is the $R^2$ value of the fit after the cut-off filter is applied. The asterisk (*) indicates which filter was chosen for each season’s final fit.

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(“online” 0.21 nm, “offline” 0.20 nm), such that the online signal encompasses several water-vapour lines instead of a single line. This is why it is called a “broad-band” DIAL. Water-vapour profiles are measured by calculating the difference in atmospheric absorption between the two signals; the difference in atmospheric absorption can be calculated by taking the ratio of the online and offline signals \((P_{\text{on}}, P_{\text{off}})\), Equation 1:

\[
\frac{P_{\text{on}}}{P_{\text{off}}} \propto \frac{\int_{-\infty}^{\infty} S_{\text{on}}(v - \nu_{\text{on}}) T_{\text{WV}}^2(v, z) \, dv}{\int_{-\infty}^{\infty} S_{\text{off}}(v - \nu_{\text{off}}) T_{\text{WV}}^2(v, z) \, dv} \tag{1}
\]

where \(S_{\text{on,off}}\) is the normalized laser spectrum for each frequency \((v)\) and \(T_{\text{WV}}(v, z)\) is the one-way atmospheric transmission due to water-vapour absorption at each frequency with altitude \((z)\). The derivation of Equation 1 can be found in Newsom et al. (2020). While Vaisala’s water-vapour retrieval algorithm is proprietary, it is similar to the broad-band water-vapour DIAL solution presented in South et al. (1998). Solving for the water-vapour transmission profile provides the number-density profile for water vapour via the Beer–Lambert law. Water-vapour mixing ratios are calculated by normalizing a standard temperature and pressure profile to temperature and pressure measurements taken by a Vaisala WXT534 present weather sensor on top of the instrument. The uncertainty in the water-vapour profile introduced by using a standard pressure and temperature profile is minimal (max of 0.2% in the water-vapour concentrations at 3 km above ground level (AGL): Newsom et al., 2020).

The technical specifications of the Vaisala preproduction DIAL can be found in Table 2. The preproduction model has been slightly modified compared with the prototype validated at the SGP, as indicated by a “*” in Table 1. It is a relatively small instrument (compared with other DIAL lidars), only slightly larger than a Vaisala ceilometer. In fact, it employs two Vaisala CL-series ceilometer telescopes for both its near and far-field measurements. The near-field telescope is optimized to retrieve water vapour and attenuated backscatter measurements from 50–400 m. The far-field telescope is optimized for measurements from 300–3,000 m. The Vaisala water-vapour retrieval algorithm uses a linear tent function to smooth the overlap region between the near and far-field measurements. The two lasers use a common optical path for both telescopes, such that the overlap functions for the online and offline signals are identical (Newsom et al., 2020). More detailed explanations of the DIAL’s design can be found in Newsom et al. (2020) and Mariani et al. (2020).

Three validation studies for the Vaisala DIAL have been published. Newsom et al. (2020) validated the prototype model against RS92 radiosondes, a Raman lidar, and an AERI instrument in the Southern Great Plains. They found that the prototype version was unbiased \((-0.01 \text{ g-kg}^{-1}\) with respect to the radiosonde and the Raman lidar. It exhibited a slight bias with respect to the AERI \((-0.22 \text{ g-kg}^{-1}\). The prototype DIAL water-vapour measurements were highly correlated (0.92–0.99) with all of the validation instruments. After acquiring a newer preproduction model, ECCC performed a validation with a semicoincident Raman lidar in Toronto, Canada (Mariani et al., 2020). Mariani et al. (2020) found that the preproduction model performed well in an urban environment and had a slight bias of 0.17 ± 0.14 g·kg⁻¹. They found that a portion of the bias was attributed to differences in geography, as the two lidars were 3.2 km apart, whereas the rest of the wet bias was due to the processing algorithm.

The DIAL accurately captured fast-moving meteorological features (e.g., a cold front) during the validation study.

Mariani et al. (2021) performed a year-long multi-instrument validation of the preproduction DIAL, including comparisons with NWP model output, after it was installed at the Iqaluit supersite. The Mariani et al. (2021) validation study was conducted over the same time period as this diurnal-cycle study. The evaluation compared the DIAL with twice-daily GRAW DFM-09 radiosondes as well as the Canadian Autonomous Arctic Aerosol Lidar

| Quantity                                      | Value                     |
|-----------------------------------------------|---------------------------|
| Laser eye safety classification                | 1M (eye-safe)             |
| Dimensions*                                   | 1.97 × 0.85 × 0.58 m     |
| Weight*                                       | 150 kg                   |
| Average time/reporting interval*              | 20-1 min⁻¹              |
| WV vertical range                             | up to 3 km               |
| Att. backscatter vertical range*              | 14.4 km                  |
| Vertical bin size*                            | 4.8 m                    |
| WV vertical resolution                        | 100–500 m                |
| Laser type                                    | Diode laser              |
| Laser wavelengths (online, offline)           | 911.0 nm, 910.6 nm       |
| Pulse energy*                                 | 9 μJ                     |
| Pulse power                                   | 25 W                     |
| FWHM pulse width                              | 220 ns                   |
| Pulse repetition rate*                        | 10 kHz                   |
| Spectral width (near/far range)*             | 0.21-0.20 nm⁻¹           |
| Telescope diameter (near/far range)*          | 150-280 mm⁻¹             |
| Receiver detector                             | Avalanche photodiode     |
| Receiver bandwidth                            | 3 MHz                    |

Note: Changes between the Newsom et al. (2020) and current preproduction versions are marked with a “*”. This is an extended version of the table presented in Mariani et al. (2020).
(CAAAL) Raman lidar. The results of the validation showed that the DIAL agrees well with the CAAAL and daily radiosondes. The campaign revealed a small yet consistent global (over the entire profile) positive bias compared with radiosondes ($0.13 \pm 0.01 \text{g kg}^{-1}$) and CAAAL ($0.18 \pm 0.02 \text{g kg}^{-1}$). This wet bias is due to an overestimation of the laser spectral width in the DIAL’s retrieval algorithm. Vaisala will be updating the spectral width in subsequent software and hardware versions. The global wet bias does not exhibit a diurnal cycle and is systematic, therefore it does not influence the diurnal-cycle calculations. The evaluation also revealed a pronounced bias and high variability in the measurements between 200 and 450 m AGL, particularly in the summer. It became apparent that the variability and bias were caused by the weighting function used to merge the near- and far-field measurements in the low signal-to-noise ratio (SNR) range gates of the near-field measurements. This issue will be addressed in following software releases. We found that this region also exhibits an internal diurnal cycle (Mariani et al., 2021). Therefore, to be conservative, we do not consider the diurnal cycles calculated between 250 and 450 m AGL in our analysis.

The DIAL produces a 20-min average water-vapour mixing ratio profile every minute (20-min running average) at 4.8-m vertical resolution. In this study, we use every 20th profile, such that we have a measurement at 00, 20, and 40 min past every hour. We also smooth the profiles to 100-m vertical resolution, where the measurement is placed in the middle of each range bin. Newsom et al. (2020) calculated the vertical resolution of the DIAL and found it to increase from roughly 100 m in the first kilometer of altitude to up to 500 m at 3 km altitude. While the bin size of our diurnal calculations is 100 m throughout the entire profile, the vertical resolution changes to 200–300 m by 1,500 m and therefore some of the bins at those altitudes could have some correlation between bins. Note that our results did not change when the resolution was decreased in height, indicating that the effect of this correlation is minimal.

### 2.4 Meteorological surface station

The two-metre surface station at Iqaluit is a standard WMO surface station. The measurements are publicly available via ECCC’s historical climate data repository (Environment and Climate Change Canada, 2020). All surface meteorological measurements (including pressure, temperature, and relative humidity) are provided hourly and we use all measurements available from January 2009–January 2020. We used measurements from 2009 onward to maintain consistency with the GNSS time series. We used the International Association for the Properties of Water and Steam (IAPWS) 1995 formula for saturation vapour pressure to convert the surface-station relative humidity measurements to water-vapour mixing ratios (Wagner and Prüß, 2002).

### 3 DIURNAL ANALYSIS METHODOLOGY

After applying the necessary quality-control filters to each data set, we divided the measurements further into seasons. Most climate and meteorological studies use a standard seasonal division such as DJF, MAM, JJA, and SON. However, this typical division is not well suited to the Arctic and subarctic due to the extreme change in daylight hours over the year, particularly for diurnal studies. Therefore, we have divided the seasons slightly differently based on the average hours of daylight in a day. We define “daylight” to be from the end of astronomical twilight in the morning to the beginning of astronomical twilight in the evening. The longest “season” is Iqaluit’s summer from April–August, with an average of 18 hr of daylight. Winter is from November–January, with an average of 8 hr of daylight. The spring (Feb and March) and fall (September and October) seasons have an average of 12.5 hr of daylight. Our seasonal division using daylight hours correlates naturally with the average amount of downward shortwave radiation received by the Arctic region. Semmler et al. (2005) calculated the average annual downward shortwave radiation cycle for Alaska and Canada above 70°N. In his calculation, the minimum amount of shortwave radiation is received from November–January. It begins to increase dramatically from February–March. The peak extends from April–July and then decreases again in August–November. We include August in the summer months, because Iqaluit is lower in latitude at 63°N than the latitudes considered in Semmler et al. (2005) and has the same amount of daylight as July.

After dividing each data set by season, we calculate the diurnal cycle for each time series. In the DIAL’s case, the diurnal cycles are calculated every 100 m in altitude. We based our diurnal-cycle calculation on the method used in Dai et al. (2002). The diurnal cycle is represented in Equations 2 and 3, where $I(t’)$ is each instrument’s time series, $I_0$ the average value for the time series, $R$ the time series residuals, and $S_n(t’)$ the diurnal cycle divided into four harmonics:

$$I(t’) = I_0 + \sum_{n=1}^{4} S_n(t’) + R. \quad (2)$$
The surface mixing ratio diurnal-cycle anomalies by season: diurnal cycles for (a) summer, (b) fall, (c) winter, and (d) spring, respectively. The black solid line is the total fitted diurnal cycle to the average surface mixing ratio hourly measurements, represented by the black crosses ("x"). The red dot-dashed line is the 24-hr cycle \(S_1\) and the blue dashed line is the 12-hr \(S_2\) cycle. Note that the magnitude of the y-axis changes with the season [Colour figure can be viewed at wileyonlinelibrary.com]

\[ S_n(t') = a_n \cos(nt') + b_n \sin(nt'). \] (3)

The amplitude \(A_n\) and phase \(\phi_n\) for each harmonic are found via \(a_n\) and \(b_n\) from Equations 4 and 5:

\[ A_n = \sqrt{a_n^2 + b_n^2}. \] (4)

\[ \phi_n = \tan^{-1} \left( \frac{a_n}{b_n} \right). \] (5)

The variable \(t'\) is the mean local solar time (LST) in radians such that \(t' = 2\pi t/24\) and \(t\) is the local solar time in hours. The first harmonic is the 24-hr cycle \(S_1\), the second is the 12-hr cycle \(S_2\), the third is the 6-hr cycle, etc. We found that the third and fourth harmonics were negligible in our analysis, therefore we will only present the magnitudes of the 24- and 12-hr cycles in this study.

4 | RESULTS

4.1 | Surface diurnal cycles

We seasonally averaged the WMO standard 2-m surface-station water-vapour mixing ratio and temperature time series and then fitted the time series to Equations 2 and 3. Figure 2 shows the seasonal diurnal cycles for the surface water-vapour mixing ratio measurements. The diurnal cycles are represented as diurnal anomalies, where the mean value of the time series has been subtracted (e.g., \(I(t') - I_0\)). The summer seasonal cycle exhibits a very large \(S_1\) (24-hr) component with an amplitude of 0.19 g·kg\(^{-1}\) (Table 3) and a maximum at 1440 LST. The \(S_2\) (12-hr) amplitude is an order of magnitude smaller at 0.01 g·kg\(^{-1}\). The fall total
TABLE 3 Surface mixing ratio diurnal-cycle amplitudes and phases for both the $S_1$ and $S_2$ cycles as a function of local solar time

| Cycle | Season | Amplitude ($g \cdot kg^{-1}$) | Phase (hr) | $I_0$ ($g \cdot kg^{-1}$) | Daily STD ($g \cdot kg^{-1}$) |
|-------|--------|-------------------------------|------------|---------------------------|-------------------------------|
| $S_1$ | Summer | 0.19                          | 2.64       | 3.69                      | 1.67                          |
| $S_1$ | Fall   | 0.08                          | 1.64       | 3.23                      | 0.99                          |
| $S_1$ | Winter | 0.02                          | 1.37       | 1.63                      | 1.02                          |
| $S_1$ | Spring | 0.05                          | 2.84       | 0.54                      | 0.41                          |
| $S_2$ | Summer | 0.01                          | -1.58      | "                         | "                             |
| $S_2$ | Fall   | 0.02                          | 0.53       | "                         | "                             |
| $S_2$ | Winter | 0.01                          | 1.96       | "                         | "                             |
| $S_2$ | Spring | 0.02                          | 1.49       | "                         | "                             |

Note: The phase is the time of maximum with respect to solar noon (e.g., 2.64 is 14.64 LST). The phase of the $S_2$ cycle is calculated with respect to the maximum closest to noon. $I_0$ is the mean value of the time series. The daily standard deviation (STD) is the standard deviation of the daily average surface mixing ratio. $I_0$ and the daily standard deviation are independent of the cycle period, therefore they are only recorded once.

The $S_1$ amplitude remains small, but does increase up to 0.02 g·kg$^{-1}$. The winter diurnal cycle begins to exhibit influence from the semidiurnal component, because the $S_1$ amplitude has decreased significantly down to 0.02 g·kg$^{-1}$, which is almost equal to the $S_2$ amplitude (0.01 g·kg$^{-1}$). The spring diurnal cycle exhibits some influence from the $S_2$ cycle, but is again largely dominated by the $S_1$ component. The spring $S_1$ amplitude increases back up to 0.05 g·kg$^{-1}$, which is slightly less than the fall amplitude. The spring $S_2$ amplitude also increases back up to 0.02 g·kg$^{-1}$, which is the same amplitude it had in the fall.

Figure 3 and Table 4 show the seasonal diurnal cycles for surface temperature. The surface temperature diurnal cycles exhibit behaviour similar to the surface mixing ratio cycles. In the summer, $S_1$ dominates with an amplitude of 1.91 °C and $S_2$ is almost non-existent at 0.04 °C. However, the fall $S_1$ amplitude is almost half that of the summer cycle at 1.08 °C. The $S_2$ amplitude increases in the fall up to 0.32 °C and is slightly visible in the total cycle. In the winter, the $S_1$ amplitude (0.32 °C) decreases and is almost

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FIGURE 3 Same as Figure 2, but for surface temperature measurements [Colour figure can be viewed at wileyonlinelibrary.com]
TABLE 4  Surface-temperature diurnal-cycle amplitudes and phases for both $S_1$ and $S_2$

| Cycle | Season | Amplitude ($^\circ$C) | Phase I ($^\circ$C) | Daily STD ($^\circ$C) |
|--------|--------|-----------------------|---------------------|----------------------|
| $S_1$  | Summer | 1.91                  | 2.85                | 0.59                 | 8.35                |
| $S_1$  | Fall   | 1.09                  | 2.25                | $-0.22$              | 4.52                |
| $S_1$  | Winter | 0.31                  | 1.81                | $-11.22$             | 8.7                 |
| $S_1$  | Spring | 1.43                  | 2.72                | $-23.65$             | 6.85                |
| $S_2$  | Summer | 0.04                  | 0.99                | "                    | "                   |
| $S_2$  | Fall   | 0.32                  | 1.40                | "                    | "                   |
| $S_2$  | Winter | 0.20                  | 1.50                | "                    | "                   |
| $S_2$  | Spring | 0.51                  | 1.31                | "                    | "                   |

Note: Same as Table 3 but for temperature.

4.2  Height-resolved diurnal water-vapour cycles

As with the surface-station measurements, the DIAL measurements were seasonally averaged over a 24-hr period. When calculating the seasonal average, it is important to consider that each lidar profile measures a different altitude range depending on the conditions of the atmosphere at the time. Vaisala uses a quality-control algorithm to determine the maximum effective range of each profile. However, Mariani et al. (2021)’s validation study revealed that it is often better to remove the 80 m below the automatic effective height, since this is where the profiles often start to diverge from the radiosonde measurements. To be conservative, we have removed the 200 m below the maximum effective height for each profile to ensure no artifacts from the retrieval algorithm affect the diurnal-cycle calculation. To determine a cut-off altitude for the seasonal average, we required that more than 50% of the time bins at any given altitude have at least 20 measurements in order to calculate a fit. This criterion created a natural cut-off altitude for each of the seasonal averages. While this cut-off is somewhat arbitrarily chosen, in the end, the $R^2$ values of the diurnal-cycle fits determine at which heights the diurnal-cycle calculations are considered statistically robust. It should also be mentioned that, due to the DIAL’s height range dependence on meteorological conditions, it is possible that the higher altitude results may be more biased towards clearer conditions during no precipitation. We will address this in Section 5.

Diurnal-cycle fits were calculated with the same method as the surface-station measurements for every 100-m altitude bin. We also calculate the “normalized” diurnal cycle, where the diurnal mixing ratio cycles are divided by the standard deviation of the daily average mixing ratio at each given altitude. The normalized cycle was first presented in Dai et al. (2002) and Wang et al. (2002) for radiosonde and microwave radiometer measurements, respectively. The normalized cycle represents the magnitude of the diurnal cycle with respect to the daily variation of water vapour. Figures 4–7 (panels a,b) show the total diurnal-cycle fits ($S_1$ and $S_2$ components added together) as both absolute and normalized diurnal anomalies, respectively, where $I_q(z)$ has been subtracted for each altitude. To highlight visually the consistency between the surface-station measurements and the lidar measurements, we plotted the surface-station total diurnal mixing ratio cycle in the first 15 m of altitude for each panel. The greyed-out sections in the first two panels are not considered in our analysis, due to the known internal diurnal cycle interference described in Section 2.3. Solid contour lines and red regions represent positive anomalies, while dashed contours and blue regions represent negative anomalies. The third panel (c) in these figures is the standard deviation of the daily average mixing ratio value at each altitude. Panel (b) is calculated by dividing the first panel by the third. Panel (d) is the $R^2$ value for each fit as a function of altitude. Panel (e) is the number of observations used to calculate each fit.

The summer diurnal water-vapour mixing ratio cycle is shown in Figure 4. The 200 m above the surface are in phase with the surface water-vapour mixing ratio cycle. The amplitude also largely stays the same between the surface and the first 200 m. Between 450 and 1,000 m, the maximum positive anomaly is still 0.1 g kg$^{-1}$, while the minimum is $-0.2$ g kg$^{-1}$. However, the maximum anomaly shifts from being centred at 1400 LST in the first 200 m to 2100 LST between 450 and 1,000 m. The $S_1$ cycle dominates below 1,000 m and appears to continue above 1 km; however, above 1,000 m the amplitude of the total cycle increases due to influence from the $S_2$ cycle. There, the maximum amplitude is 0.3 g kg$^{-1}$ at 0600 LST with another maximum equally as high at 2000 LST. The minimum occurs at 0900 LST, with a change between $-0.2$ and $-0.3$ g kg$^{-1}$. The standard deviation of the daily water-vapour mixing ratio content varies between 1.4 and 1.7 g kg$^{-1}$ per day. The summer diurnal cycle makes up roughly 5% (near the surface) and up to 15% (above 1 km)
The summer diurnal water-vapour cycle and statistics measured by the surface station (0–15 m) and the DIAL (100+ m).

From left to right: (a) absolute diurnal cycle relative to the summer average for each altitude in units of water-vapour mixing ratio (g kg⁻¹) as a function of local solar time, the grey region is considered unreliable due to known instrumental effects; (b) the normalized diurnal cycle with respect to the standard deviation of the daily average over the summer; (c) the standard deviation of the daily average of the water-vapour mixing ratio over the summer months (dividing a by c produces b); (d) the $R^2$ of the diurnal-cycle fits as a function of altitude; (e) the number of observations used to calculate the diurnal-cycle solutions in (a). Note that the surface measurements are hourly resolution and the DIAL at 20-min resolution [Colour figure can be viewed at wileyonlinelibrary.com]

of the daily variability. However, it is important to note that the quality of the diurnal fits decreases above 1,000 m due to the decrease in the number of available observations and should be treated with caution. There are fewer profiles available above 1,000 m and increased noise, considering that only 1 year of summer measurements was available. Fits with $R^2$ values below 0.5 are generally considered untrustworthy.

The fall diurnal mixing ratio cycle is presented in Figure 5. Similarly to the summer, the phase and amplitude in the first 200 m agrees well with the surface measurements. The cycle maximum occurs at 1100 LST below 250 m and has a maximum between 0.10 and 0.15 g kg⁻¹. The minimum anomaly is ~0.05 g kg⁻¹ at 0300 LST and is much smaller than the summer’s minimum at the same altitude range. Between 450 and 1,000 m, the phase and amplitude are largely consistent except for slightly larger negative anomalies above 800 m. Above 1,000 m, the phase of the cycle shifts by 12 hr. Throughout the entire altitude range, the $S_1$ cycle remains dominant. The standard deviation of the daily mean water-vapour mixing ratio reduces down to a range of 0.8–1.0 g kg⁻¹. Similarly to the normalized cycles in the summer, below 1,000 m the diurnal cycle is less than 10% of the magnitude of daily variability. However, above 1,000 m the diurnal cycle is much larger—up to 25% of the daily variability’s magnitude. While the diurnal-cycle amplitudes remained close to the summer amplitudes, the daily variability decreased by 38%, thereby increasing the contribution of the diurnal cycle. As with the summer cycle, it is important to consider the quality of the fits at each altitude. Above 600 m, $R^2$ values drop below 0.5. It is interesting to note that, despite the relatively small decrease in the number of observations from the surface, the solution at 750 m has a very low $R^2$ value. While low $R^2$ values are due to poor fitting, the reason for the poor fits changes when there are still a relatively large number of points. In this case, it could indicate a region where there is no diurnal cycle present with which to calculate a fit.
Figure 6 is the winter diurnal cycle. The amplitude of the diurnal cycle over all altitudes drops by an order of magnitude. With amplitudes so small, it is difficult to say whether or not the DIAL detects a diurnal signal in the winter. Despite similar amounts of observations to the fall, the $R^2$ values are still quite poor: less than 0.5 for all altitudes except at 750 m. Below 1,000 m, the diurnal anomalies range between $\pm 0.04$ g·kg$^{-1}$. Below 1,000 m, the DIAL measures a maximum in water vapour at 1300 LST, which is consistent with the surface-station maximum. The surface-station mixing ratio cycle includes a larger component from the $S_2$ cycle which does not appear in the DIAL’s measurements. Above 1,000 m the phase of the total cycle shifts slightly to 1100 LST and the amplitude of the anomalies doubles to $\pm 0.08$ g·kg$^{-1}$. The $S_1$ cycle is more apparent than the $S_2$ cycle at higher altitudes as well. The variability of the water vapour decreases down to an average of 0.5 g·kg$^{-1}$. However, despite the decrease in daily variability, the magnitude of the normalized diurnal cycle decreases down to 0–5% of the daily variability below 1,000 m and remains between 5 and 25% above. The small relative amplitudes below 1,000 m would continue to suggest that the diurnal cycle is not the main source of daily variability.

The spring (Figure 7) has amplitudes similar in magnitude to winter, except at the surface, which is wetter than the first DIAL bins by about 0.02 g·kg$^{-1}$. However, the surface and low-altitude DIAL measurements agree in phase with maxima at 1300 LST. Below 200 m, the diurnal cycle is between 5 and 15% of the daily variability. The normalized amplitudes in the spring are slightly larger than in winter because the daily variability of the water decreases from 0.5 to 0.25 g·kg$^{-1}$, while the diurnal-cycle amplitudes remain relatively constant. There is an abrupt 12-hr shift in phase at 500 m and a decrease in amplitude of 0.02 g·kg$^{-1}$. At 1 km, the $S_2$ component begins to influence the total cycle such that we see a maximum at 0700 and 2200 LST. As with other seasons, we do see an increase in amplitude with altitude, particularly in the enhanced maximum at 1 km of 0.1 g·kg$^{-1}$ (up to 20% of the daily variability). Spring has slightly fewer measurements available than winter, as it uses four months of observations instead of six. As with winter, $R^2$ values for spring are extremely low, therefore results for spring must be treated with caution.

Figures 4–7 showed the summation of the diurnal and semidiurnal cycles, or the total diurnal cycle. Figure 8 separates the $S_1$ and $S_2$ amplitudes and phases and shows their seasonal variations. In the summer, $S_1$ amplitudes are
larger than $S_2$ amplitudes by roughly a factor of 5 below 1 km and a factor of 2–3 above. In the fall, the $S_1$ amplitudes are generally twice the semidiurnal amplitudes. However, in the winter and spring, the $S_2$ amplitudes are generally equal to or even larger than the $S_1$ amplitudes. Both the $S_1$ and $S_2$ cycle amplitudes exhibit seasonal variation. However, it should be noted that, due to the short time series available from the DIAL, any seasonal variation should be treated with caution, particularly in the winter and spring, where $R^2$ values are low. The $S_1$ cycle has a strong seasonal variation from the surface to 1.5 km, with larger amplitudes in the summer that decrease in the winter. The spring amplitudes do not increase significantly above the winter amplitudes, except at 50 m and between 1,000 and 1,250 m. Unlike the $S_1$ cycle, the $S_2$ cycle amplitudes do not always change with the season. Below 250 m and between 750 and 1,000 m there is very little seasonal change apparent; however, there is some seasonal variation between 250 and 750 m, which shows similar behaviour to the diurnal amplitudes. For example, at those altitudes, summer amplitudes are the largest, with winter and spring sharing the minimum amplitudes. Both diurnal and semidiurnal amplitudes increase with altitude.

The phases in Figure 8 are the time of the maximum shown with respect to solar noon. The $S_1$ phases do exhibit some seasonal variation between 100 and 1,500 m. Summer maxima occur 4 hr after solar noon closer to the surface and up to 10 hr after solar noon around 750 m. Fall maxima occur closer to noon from 100 m–1 km where they suddenly shift to 2300 LST, as was seen in Figure 5. However, it is important to keep in mind that the $R^2$ values at these altitudes are less than 0.5. The winter diurnal maxima occur before solar noon below 250 m, but shift by 12 hr to 1700 LST between 500 and 1,000 m. After 1,000 m, the maxima gradually shift back to solar noon. The spring maxima, on average, occur between 1200 and 1400 LST below 250 m and 2300–0400 LST between 500 m and 1 km. Above 1200 m they occur around 1130 LST. The $S_2$ phases vary seasonally, except below 250 m and above 1 km. Below 250 m, the summer, fall, and spring maxima occur within 2 hr of each other between 1100 and 1300 LST. The winter maximum occurs later at 1500 LST. Between 500 and 1,000 m, there are small changes in the phase of the order of 1–2 hr around solar noon. The summer maxima at those altitudes are generally 2.5 hr later than those observed in winter, spring, and fall. Above 1 km, there is a larger variation in the $S_2$ phase with season. The summer maxima
again occur later in the day at 1800 LST, fall maxima occur at solar noon, and spring maxima at 0930 LST. Interestingly, the winter semi-diurnal maxima around 1 km occur at 2100 LST but then shift very quickly back to solar noon from 1,200–1,800 m, but given the low winter $R^2$ values this may not be reliable.

4.3 Total column diurnal WV cycles

After averaging the GNSS time series seasonally, we fit the IWV series to Equations 2 and 3. The results of the fits for each season, along with their $R^2$ values, are presented in Figure 9. The amplitudes and phase for each fit are shown in Table 5. The same diurnal fit analysis was applied to the GNSS measurements. The first cycle shown in Figure 9a is the summer cycle from April–August. There is a large $S_1$ component present with an amplitude of 0.10 mm. Interestingly, the semi-diurnal cycle is also significant in the summer cycle, which creates an extended maximum from 1200–0100 LST. The summer maximum for the total cycle occurs at 1500 LST and the minimum occurs at 0730 LST. Figure 9b is the fall diurnal cycle. While the amplitude of the 24-hr period remains the same, its phase is shifted such that the $S_1$ maximum occurs at the same time as the $S_2$ maximum. This creates a total maximum in integrated water vapour at 0200 LST. The $S_2$ amplitude increases up to 0.11 mm. The strength of the 12-hr cycle creates two minima in the total cycle at 0900 and 1900 LST. Figure 9c is the winter cycle, which shows a prominent 12-hr cycle due to a weakening in the $S_1$ amplitude down to 0.04 mm. The $S_2$ amplitude also decreases down to 0.05 mm. Due to the influence of the $S_2$ cycle, the total cycle has two maxima at 0100 and 1500 LST and minima at 0800 and 1930 LST. The largest change in water vapour is overnight, as both the 24- and 12-hr cycles are decreasing during that time. The last subfigure is the spring cycle. The spring cycle changes very little with respect to the winter cycle. The $S_1$ amplitude increases slightly up to 0.05 mm, while the $S_2$ amplitude remains the same as it was in the winter. However, the phase of the $S_1$ cycle is shifted from a maximum at 2200 LST in the winter to 1900 LST in the spring. This creates an overall maximum at 1500 LST instead of at 0100 LST.

Generally, we see that the amplitude of the $S_1$ cycle is at a maximum during the summer months and decreases in amplitude into the winter and spring. With the exception of the increase in amplitude during the fall, the $S_2$ cycle...
remains at the same amplitude throughout the rest of the year. The fall cycle exhibits a large jump in water vapour between 2300 and 0000 LST, which is also observed in the surface-station mixing ratio measurements (Figure 2). This is due to the steep decrease in IWV over September and October (Figure 1). Similar decreases or increases are not apparent in the other seasons, due to either relatively small changes over the seasons (e.g., winter and spring) or the season encompassing both positive and negative changes in water vapour (e.g., summer).

5 | DISCUSSION

As mentioned previously, very few diurnal analyses of water vapour have been conducted in the Arctic, and none in the Canadian Arctic. However, it is possible to compare our work with studies at similar latitudes, as those locations will receive similar amounts of solar radiation over the course of the day and year. Some differences between studies should be expected, due to differences in local geography and climate. Dai and Trenberth (2004) conducted a global study of diurnal surface-temperature observations. They found that, in the winter (DJF), near Iqaluit the amplitude of the diurnal cycle was roughly $0.5^\circ$ C, which agrees well with our amplitude of $0.3^\circ$ C. In the summer (JJA), they found amplitudes of around $1.0^\circ$ C, which is slightly smaller than our $1.9^\circ$ C. However, their contour resolution is coarse and the time period for their study is 1976–1997. The bias between our two studies could be due to climate change, the difference in months used for our average, or the difference in resolutions.

Dai et al. (2002) (henceforth D2002) calculated diurnal IWV cycles in Alaska between 62.5° and 67.5°N latitude. They found summer (JJA) $S_1$ amplitudes between 0.12 and 0.27 mm and $S_2$ amplitudes between 0.55 and

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**Figure 8** The amplitudes and phases of the water-vapour diurnal cycle calculated by the DIAL for each season. Phases are shown as the hour of maximum with respect to solar noon (same as previous tables, LST). (a) $S_1$ amplitudes, (b) $S_1$ phases, (c) $S_2$ amplitudes, and (d) $S_2$ phases [Colour figure can be viewed at wileyonlinelibrary.com]
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FIGURE 9  Same as Figure 2 but for GNSS IWV measurements. Note the change in y-axis for each season [Colour figure can be viewed at wileyonlinelibrary.com]

TABLE 5  GNSS diurnal-cycle amplitudes and phases for both the $S_1$ and $S_2$ cycles

| Cycle | Season     | Amplitude (mm) | Phase   | $I_0$ (mm) | Daily STD (mm) |
|-------|------------|----------------|---------|------------|----------------|
| $S_1$ | Summer     | 0.10           | 5.93    | 10.36      | 3.86           |
| $S_1$ | Fall       | 0.10           | −9.85   | 8.38       | 3.06           |
| $S_1$ | Winter     | 0.04           | 8.92    | 3.70       | 0.74           |
| $S_1$ | Spring     | 0.04           | 7.36    | 3.91       | 1.33           |
| $S_2$ | Summer     | 0.04           | 2.28    | "          | "              |
| $S_2$ | Fall       | 0.11           | 2.24    | "          | "              |
| $S_2$ | Winter     | 0.04           | 2.39    | "          | "              |
| $S_2$ | Spring     | 0.05           | 1.60    | "          | "              |

0.8 mm. While our $S_1$ amplitudes agree well, our $S_2$ amplitudes are an order of magnitude smaller. Jakobson et al. (2009) (henceforth J2009) and Jakobson et al. (2014) (henceforth J2014) calculated diurnal IWV cycles for the Baltic Sea region between 53° and 68°N. J2009 used GNSS measurements, while J2014 used the BaltAN65+ reanalysis model. Jakobson et al. (2009) found summer (JJA) IWV amplitudes between 0.1 and 0.8 mm. Iqaluit’s amplitudes (0.1 mm in the summer) are on the lower end of their distribution; however, they do agree. Due to the large geographical coverage and variability of the J2009 study, a large amount of variation in the diurnal cycle is to be expected. We generally agree in phase with J2009, with their average minimum occurring at 0500 LT and 0700 LT and ours at 0730 LST; our maxima also agree with theirs between 1500 and 1700 LT and ours at 1600 LST. The lower amplitude we observe in the summer could be due to the inclusion of March and April in our summer months, or simply differences in climate. We also observe similar amplitudes in the winter (DJF) (∼0.01 mm) and an increase in contribution from the $S_2$ component. In the fall (SON), J2019 has a slightly larger amplitude (0.16 mm) than ours (0.10 mm) with a maximum occurring overnight and a minimum close to noon; we also observe a phase with a maximum overnight around 0200 LST and a minimum around 1400 LST. Our spring amplitudes are much smaller than J2009’s, most likely due to the difference in
our definitions of spring (MAM, J2009 versus FM). Their spring is much later in the year than ours and therefore receives more solar radiation, which results in larger diurnal amplitudes (0.25 mm versus 0.04 mm). They also observe a much larger $S_1$ component than we do (based on their figure, not a harmonic analysis), again likely due to the difference in our spring definitions. Despite the difference in spring definitions, we find that our GNSS cycles are in good agreement in both phase and magnitude with the J2009 study. The J2014 reanalysis agrees with the J2009 observations, therefore we are in good agreement with them as well.

The agreement between the D2002, J2009, and J2014 studies suggests that the quality-control filter for the GNSS did not affect our results and that our GNSS cycles are within the expected amplitude and phase ranges. One interesting difference between the three studies is the difference in the $S_2$ component amplitude. J2009 and J2014 observe very little influence in the $S_2$ component between March and August. However, the $S_2$ component is more apparent from September–February, when the $S_1$ component decreases. In D2002, the $S_2$ component is the larger component in the summer and smaller in the winter. This suggests that the $S_2$ component is likely not correlated with latitude and the solar radiation annual cycle and is influenced by some other driver. More research on the differences in climate and geography between the locations is required to determine possible drivers.

J2014 also calculated height-resolved specific humidity diurnal cycles of the Baltic Sea region using the BaltAN65+ reanalysis model from 1,000–700 hPa (figure 5 in J2014). Comparisons of the height-resolved analysis reveal similar agreement between the two studies, with the exception of the measurements at 1,000 hPa. At 1,000 hPa (~100 m in Iqaluit), J2014 measures a maximum at 0100 LT and minimum at 1300 LT, while we observe the opposite. However, despite the difference in phase, we both observe amplitudes of between 0.1 and 0.2 g·kg⁻¹. The phase difference could be due to the difference in months used for the calculation or in diurnal drivers at the surface. For example, J2014 states that radiative fog in the early morning is a frequent phenomenon which contributes to higher atmospheric water-vapour concentrations; radiative fog is infrequent in Iqaluit (Hanesiak and Wang, 2005). The timing of precipitation is also an important factor. Dai et al. (2002) observed that precipitation occurring in the afternoon and evening would shift the diurnal water-vapour maximum to later hours/evening and cause a decrease in atmospheric water vapour in the afternoon. At 950 hPa (~500 m), J2014 observes negative anomalies between −0.05 and −0.2 g·kg⁻¹ at 0700 LT and positive anomalies at 1300 LT and 1900 LT between 0.01 and 0.1 g·kg⁻¹. This agrees with our minimum at 0600 LST of −0.15 g·kg⁻¹ and maximum at 1600 LST of 0.15 g·kg⁻¹. At 900 hPa (~1 km) the phases remain the same, but the magnitudes increase in both sets of observations and agree with each other. Finally, at 850 hPa (~1.5 km), while we see a positive anomaly at 0000 LST, J2014 does not. However, we both measure strong minima around 0600–0800 LT of −0.2 g·kg⁻¹ or lower (J2014), and −0.3 g·kg⁻¹. At 1300 LT, J2014 measures generally weaker positive anomalies than ours at 850 hPa between 0 and 0.1 g·kg⁻¹; we measure positive anomalies between 0.05 and 0.1 g·kg⁻¹. Finally, at 1900 LT J2014 measures maxima of 0.1 g·kg⁻¹ or greater, while we measure maxima of 0.2–0.3 g·kg⁻¹ at 1900 LST. Thus, even though our $R^2$ values are less than 0.5 above 1 km, it appears that the DIAL measurements may be capturing the diurnal cycle fairly well, at least for the summer months. The good agreement with J2014 at the higher altitudes also suggests that any meteorological bias in our results should be minimal. While this provides increased confidence in our results for other seasons with $R^2$ values around 0.5, they must still be treated with caution, as we do not have any studies for comparison.

Now that we have established confidence in the diurnal cycles, we can discuss the dynamics and possible drivers influencing the different seasonal cycles at various altitudes. Given the diurnal cycle’s dependence on evapotranspiration, we would expect to see correlation between the surface temperature and the water-vapour cycles. In the summer, the temperature cycle has a strong $S_1$ component with an amplitude of 2 °C with a maximum at 1430 LST. The total surface mixing ratio cycle also exhibits a strong $S_1$ component and a very small $S_2$ component. The surface mixing ratio maximum slightly precedes the temperature maximum by roughly 15 min. However, since our temporal resolution is hourly, the difference of 15 min is not significant. The IWV cycle, on the other hand, has a clear lag in the $S_1$ component of roughly 3 hr with respect to the surface temperature maximum, thus creating a 2-hr lag in the total IWV cycle with respect to the surface temperature maximum. To understand the lag in the IWV, we can relate the total column to the height-resolved measurements. The first 250 m of the DIAL measurements are in phase with the surface-station mixing ratio measurements and agree in magnitude (Figure 4). The DIAL measures a shift in phase of 4–5 hr from the surface above 500 m, which likely contributes to the later maximum in the IWV $S_1$ cycle. Such a shift in phase with altitude is expected when water vapour is transferred from the surface via convection.

The phase shift between 500 and 1,000 m could be due to multiple processes, including the transport of moisture via wind or the evolution between boundary layer and free troposphere dynamics. Iqaluit is a coastal town
that experiences sea breezes in the later half of the summer, which are sources of moisture (Mariani et al., 2018). Sea-breeze dynamics could enhance water-vapour concentrations in the afternoon, as extra moisture from the ocean is brought over land. J2014 found that the sea-breeze effect off the Baltic Sea was limited to between 1,000 hPa (surface) and 900 hPa (∼1,000 m), with the strongest winds at 950 hPa (around 500 m in Iqaluit during the summer). While it is possible that the Iqaluit summer cycle is influenced by sea breezes, it is unlikely they are the main cause, since they only occur in July and August. Height-resolved diurnal wind cycles are necessary to determine their contribution to the phase changes observed in the DIAL. D2002 states that the diurnal cycle of large-scale vertical motion can create negative anomalies in the free troposphere during the day and positive anomalies at night. Conversely, the boundary-layer diurnal cycle is driven by evapotranspiration and latent heat, which creates increases in water vapour during the day and decreases at night, which we observe in the first 200 m. The average summer boundary-layer height for the Canadian Arctic is around 600 m (Esau and Sorokina, 2010). The small contribution (5%) of the diurnal cycle to the overall variability in the altitude region between 500 and 800 m would suggest that the water-vapour variability is more influenced by weather on nondiurnal scales, such as mesoscale systems. This is what we would expect of summer weather in Iqaluit in the region around the boundary-layer height. Thus, the phase shift between 500 and 1,000 m could also be the transition between the boundary-layer diurnal cycle and the free troposphere cycle.

Above 1 km, we observe positive anomalies from 1200 LST to 0600 LST. The maximum in water vapour is extended due to a shift in phase from the $S_2$ component. The DIAL summer (Figure 4) and spring (Figure 7) total diurnal cycles show visible contributions from the semidiurnal period above 1 km, which are not seen in the other seasons. However, the semidiurnal amplitudes in the fall and spring are just as large as those in the summer and winter seasons, respectively (Figure 8). The semidiurnal cycles are visible in the total cycles in the summer and spring, due to their phase shift with respect to the $S_1$ cycle. In the summer and spring, the $S_1$ and $S_2$ hour maxima are shifted such that the $S_2$ maxima occur on the inflection points of the $S_1$ cycle. This creates a more visible impact on the total cycle, such as the extended maximum in the summer and the second maximum in the spring. In contrast, in the winter and fall, the $S_1$ and $S_2$ hour cycles are in phase, such that in a 24-hr period the maxima of the $S_1$ and $S_2$ cycles occur almost simultaneously as well as at the minimum of the $S_1$ cycle. The minimum of the $S_1$ cycle then cancels, or greatly minimizes, the maximum of the $S_2$ cycle, such that we observe close to zero change in water vapour. The destructive interference then appears to enhance the 24-hr component.

In the fall, the surface mixing ratio, temperature, and DIAL mixing ratio cycles up to 1 km are in phase with maxima at approximately 1300 LST and agree in amplitude as well (Figure 5). Unlike the summer cycle, there is no shift in phase between 500 and 1,000 m and the cycle remains the same from the surface up to 1,000 m. Instead, at 1,000 m there is an abrupt shift in the $S_1$ phase (Figure 8b). However, the semidiurnal phase does not change (Figure 8d). If this phase shift is due to the transition between boundary layer and free-troposphere dynamics, we would expect it to occur lower in the atmosphere. The average boundary-layer height in September–October is between 300 and 400 m (Esau and Sorokina, 2010). However, as mentioned in the previous section, low $R^2$ values, combined with relatively large numbers of points to inform the fit, could suggest that the model used for the fit was not appropriate at this height. There may indeed be little to no diurnal cycle at all between 700 and 1,000 m. The lack of diurnal cycle could suggest that this is a transition region.

While the summer results and comparisons with J2014 suggest that we may be justified in accepting results with lower $R^2$ values, without a comparison study it is not possible to make conclusions at those heights with complete confidence. However, we can compare with the GNSS instrument to examine consistencies between the two instruments. The fall GNSS cycle exhibits a very strong $S_2$ component with maxima at 0300 and 1400 LST, which coincide with the maxima observed by the DIAL in the first 1 km and above 1 km, respectively. It is therefore possible that the $S_2$ cycle observed by the GNSS could partially be due to the superposition of the two DIAL $S_1$ phases in their respective height regimes. The DIAL $S_2$ phase is almost constant with altitude, with maxima between 1000 and 1200 LST (and again between 2200 and 0000 LST), which is about 2 hr earlier than the GNSS $S_2$ maxima. The shift between the instruments’ $S_2$ phases could be due to the difference in dates used in the two instruments’ time series. Limiting the GNSS measurements to the same dates as the DIAL improved the agreement and shifted the GNSS $S_2$ phase 1 hr earlier.

The winter and spring total diurnal cycles across all instruments and variables (water-vapour mixing ratio and temperature) are impacted more by the semidiurnal cycles than the other seasons. This is mostly caused by the decrease in $S_1$ amplitude due to the decrease in incident solar radiation at the surface, and not necessarily due to an increase in the semidiurnal amplitude.
(Figure 8, Tables 3, 4). Only the surface temperature $S_2$ amplitude increases in the spring. Uncertainties in the measurements are higher in the winter and spring, due to lower water-vapour concentrations and a higher frequency of precipitation events. While both winter and spring show good agreement in magnitude between the DIAL and surface-station mixing ratio measurements, they do differ slightly in phase in the winter. As in the summer and fall, the DIAL observes a shift in the $S_1$ phase above 500 m, which also coincides with the GNSS $S_2$ maxima in both the winter and spring. The DIAL $S_2$ phase does not agree with the GNSS $S_2$ phase in either season and the disagreement is not caused by a difference in the time series. The differences in the $S_2$ components of the two instruments are due either to the high uncertainties in the DIAL measurements or an additional contribution from the semi-diurnal component above the DIAL’s observable range.

The semi-diurnal component is extremely difficult to measure due to its small amplitude. At the surface, it represents a 0.2–3% change from the daily average (depending on the season), which requires a long timeline to discern with any significance. It makes up a similar percentage of the IWV cycle. The semi-diurnal component of the diurnal water-vapour cycle has been observed globally in IWV cycles (Dai et al., 2002; Galisteo et al., 2011; Hocke et al., 2017; Torri et al., 2019), but its drivers are not often discussed. In our comparisons between D2002 and the Jakobson studies, we saw that the $S_1$ component is relatively consistent across latitude, but the $S_2$ component is not. More global studies are needed to determine if this is true. We found that the $S_2$ Iqaluit surface water-vapour mixing ratio and IWV $S_2$ amplitudes were mostly constant throughout the year (Table 3). At most altitudes measured by the DIAL, the water vapour $S_2$ amplitude is at a maximum in the summer and fall and a minimum in the winter and spring. However, due to the low $R^2$ values in the winter and spring it is difficult to make any concrete seasonal variation conclusions with the DIAL measurements.

Studies of mesospheric diurnal water-vapour cycles found that the semi-diurnal component at those altitudes was likely driven by advection (Haefele et al., 2008; Hallgren and Hartogh, 2012). However, in the mesosphere water vapour is considered an atmospheric tracer, due to its long lifetime and lack of a source on the diurnal scale. Such is not the case in the troposphere. An in-depth analysis of the diurnal wind cycle, in addition to surface and atmospheric water-vapour flux rates at Iqaluit, is required to determine how much of the diurnal and semi-diurnal water-vapour cycle at Iqaluit is driven by advection, but that is beyond the scope of this article.

6 SUMMARY AND CONCLUSIONS

The diurnal atmospheric water-vapour cycle is a critical component of the hydrological cycle. No diurnal water-vapour studies of the Canadian Arctic have been conducted, due to its remoteness and harsh environment for instruments. Radiosondes typically do not provide enough temporal coverage to observe rapid variations or the semi-diurnal component. Additionally, radiosondes are subject to an instrumental diurnal cycle due to radiation bias, and therefore should not be used for diurnal-cycle analysis unless the bias is corrected (Dirksen et al., 2014). Operational and automatic instruments with hourly measurements are better suited to capturing the diurnal cycle. Lidars and microwave radiometers are ideally suited to fill this gap. We have successfully quantified diurnal water-vapour and temperature cycles using surface meteorological station, the new preproduction Vaisala DIAL, and GNSS measurements for Iqaluit, Nunavut. We used DIAL measurements from September 2018–June 2020 and surface-station and GNSS measurements from January 2009–June 2020. We divided the analysis into four seasons: summer (April–August), fall (September–October), winter (November–January), and spring (February–March). This particular seasonal division was chosen based on the amount of solar radiation received at the surface in Iqaluit and the number of daylight hours (Section 3). We then used harmonic analysis to determine the diurnal ($S_1$, 24 hr) and semi-diurnal ($S_2$, 12 hr) components of the diurnal cycle for each instrument’s time series. In the DIAL’s case, the harmonic analysis was applied to each 100-m altitude bin.

We compared our measurements with three studies that also calculated surface station, IWV, and height-resolved water-vapour diurnal cycles. Dai and Trenberth (2004) measured diurnal temperature cycles globally and we found them to be in good agreement with ours in the winter, but our amplitudes are slightly larger in the summer by 0.09 °C. The difference in the summer is likely due to a difference in resolution between the two studies, but could also be due to climate change, as Dai and Trenberth (2004) used measurements between 1976 and 1997. We compared the surface water-vapour mixing ratio cycles with those in J2014 and, while our amplitudes agree, the phases are 12 hr out of phase with their 1,000-hPa calculations. J2014 observed radiative fog overnight, which increases the amount of water vapour in the first 100 hPa; this was not observed in Iqaluit. However, our surface water-vapour mixing ratio cycles agreed well (within 0.01 g·kg$^{-1}$) with our DIAL measurements; therefore, the difference in phase between the two studies is likely due to a difference in geography, climate, or
local meteorological features such as radiative fog. Comparisons of the Iqaluit summer height-resolved diurnal water-vapour cycle with the J2014 height-resolved cycles for the Baltic Sea region showed good agreement between 950 and 850 hPa (500–1,500 m) in both magnitude and phase. This is encouraging, given that the two studies agree into the range where the DIAL $R^2$ values are less than 0.5. Their agreement would suggest that, even with relatively high noise in the time series, the DIAL is still able to resolve the general diurnal cycle.

We compared the IWV diurnal cycles from the GNSS with Jakobson et al. (2009)’s Baltic GNSS cycles. We found that the Iqaluit amplitudes in the summer, winter, and fall were within the range of amplitudes they calculated for the Baltic Sea region. The Iqaluit and Baltic GNSS cycles also agreed in phase in the summer, winter, and fall within 1–2 hr. The Iqaluit spring amplitudes and phases were significantly different; however, we believe this is due to the different definitions of spring used in each study. We also verified our IWV results against the Alaskan diurnal IWV cycles in Dai et al. (2002). We found that the Iqaluit summer $S_1$ amplitude (0.1 mm) was close to the range of amplitudes presented in Dai et al. (2002) (0.12–0.27 mm). However, the Iqaluit $S_2$ component is an order of magnitude smaller than that of the Baltic (but comparable with Jakobson et al., 2014). Therefore, this would suggest that the $S_1$ amplitude is correlated with latitude, but the $S_2$ amplitude and phase may not be. A larger global study is needed to confirm the relationship between the $S_2$ amplitude and latitude, as well as a deeper analysis into the drivers of the $S_2$ component.

We then examined the relationships between the surface, profile, and total column cycles. Across all seasons, except for a slight phase shift in the DIAL’s winter measurements, we found excellent agreement in phase and magnitude (differences less than 0.02 g·kg$^{-1}$) between the surface measurements and the first 250 m of the DIAL measurements. The surface and first 500 m of altitude are dominated by the $S_1$ component, which is expected in the boundary layer, where evapotranspiration is the primary driver. Above 500 m, the diurnal cycles shift in phase to maxima occurring later in the day, except in the fall, where the diurnal cycle is constant in phase up to 1 km. Above 1 km, all seasons except spring observe an almost 12 hr shift in the diurnal component. In spring, the shift occurs sooner at 500 m; however, low $R^2$ values in the winter and spring require that those results be treated with caution. The shift in phase is expected when transitioning between boundary-layer and free-troposphere dynamics.

The largest water-vapour mixing ratio amplitudes are observed in the summer, with amplitudes between 0.1 and 0.3 g·kg$^{-1}$ depending on the altitude. In the winter, the amplitudes are an order of magnitude smaller, between 0.02 and 0.08 g·kg$^{-1}$. In general, the amplitude of the diurnal cycle (both $S_1$ and $S_2$ components) increases with altitude for all seasons (Figure 8), although the $S_2$ amplitudes increase at a slightly faster rate. While the DIAL $S_1$ amplitudes are correlated with the solar annual cycle, the $S_2$ component is not; the surface water-vapour mixing ratio $S_2$ amplitude is also constant throughout the year. This, in addition to the IWV $S_2$ component’s lack of correlation with latitude, may suggest that the $S_2$ component is not driven by incident solar radiation.

The GNSS IWV $S_1$ amplitude experiences a strong seasonal cycle with a maximum of 0.10 mm in the summer and a minimum of 0.04 mm in the winter. The GNSS observes IWV cycles with a very strong semidiurnal component in all seasons. The semidiurnal component is not as visible in the total cycles of the surface station and the DIAL. The DIAL and the GNSS measurements generally agree in phase, with maxima generally occurring within 1–2 hr of each other, except in the winter. When comparing the DIAL and GNSS measurements, we found that the maxima of the GNSS semidiurnal cycle ($S_2$) coincided with the maxima of the diurnal ($S_1$) components of the DIAL, but not always with the semidiurnal component of the DIAL. This raises the question: how much of the semidiurnal component in the IWV is due to the superposition of the diurnal component throughout the troposphere? In the boundary layer, the diurnal cycle is largely driven via evapotranspiration, which creates positive anomalies during the day, whereas in the free troposphere the diurnal cycle is influenced by large-scale vertical motion or convection, which creates positive anomalies during the night and negative anomalies during the day. The superposition of the two cycles in the boundary layer and free troposphere can create a semidiurnal component in the IWV cycle. We also observed that the semidiurnal water-vapour amplitude increases with altitude. It is also possible the GNSS semidiurnal component is created by contributions from altitudes that are not observed by the DIAL.

We have shown that the DIAL can calculate diurnal water-vapour cycles accurately using 2 years of measurements. However, more measurements will be required to resolve the winter and spring cycles better with altitude. Diurnal water-vapour cycles are driven by multiple processes, including evapotranspiration, precipitation, convection, and winds, making them excellent tools for NWP model verification and validation. While GNSS measurements are now standard inputs for NWP models, lidar measurements are not. Therefore, these height-resolved DIAL diurnal cycles are an excellent source for independent verification. Future work will involve case studies evaluating the ECCC NWP model’s performance during rapid changes in IWV observed during the fall and winter. We have also shown that, while IWV and surface diurnal
cycles are important measurements that provide valuable insight into the hydrological cycle, height-resolved measurements can add even more. The DIAL measurements have provided a new, detailed insight into the dynamics in the water-vapour cycle in Iqaluit, particularly with regards to the nature of the $S_2$ component. Height-resolved horizontal wind cycles as well as water-vapour fluxes and vertical wind measurements, in addition to the DIAL water-vapour profiles, can provide a more complete view of the hydrological cycle at Iqaluit. This study illustrates the need for high temporal and vertical resolution measurements to characterize the hydrological cycle better within the boundary layer and free troposphere.

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AUTHOR CONTRIBUTIONS
Shannon Hicks–Jalali: data curation; project administration; resources; supervision; writing - review editing.
Zen Mariani: data curation; project administration; resources; supervision; writing – review and editing.
Robert W. Crawford: data curation; resources; writing – review and editing.

CONFLICT OF INTEREST
The instrument manufacturer had no role in the design of the study; in the collection, analyses, or interpretation of data used to perform the instrument or model comparisons; in the writing of the manuscript, or in the decision to publish the results.

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