An Attempt to Retrieve Continuous Water Vapor Profiles in Marine Lower Troposphere Using Shipboard Raman/Mie Lidar System

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Abstract

A shipboard lidar system was examined the capability to retrieve detailed variation of the water vapor mixing ratio in and above the marine atmospheric boundary layer (MABL). The water vapor mixing ratio is retrieved from the ratio of Raman lidar signals by water vapor and by nitrogen, with the help of radiosonde data beside. Data obtained during two special observations, Pre-YMC and YMC-Sumatra, off the west coast of Sumatra island were examined.

The mixing ratio was retrieved in the nighttime over 1 km height with the resolution of 10-minutes in temporal and 120-meters in vertical. The root mean square difference from the radiosonde data is about or less than 1 g/kg in MABL. A case study demonstrates that the retrieved spatiotemporal variation of water vapor mixing ratio captures meso-scale drying and moistening in detail. The capabilities of the retrieved data were well demonstrated, while number of improvements are expected in future work.

(Citation: Katsumata, M., K. Taniguchi, and T. Nishizawa, 2020: An attempt to retrieve continuous water vapor profiles in marine lower troposphere using shipboard Raman/Mie lidar system. SOLA, 16A, 6–11, doi:10.2151/sola.16A-002.)

1. Introduction

The water vapor in the lower troposphere is a key parameter in the energy and the hydrological cycle. Especially over the ocean, evaporation from the ocean surface to the marine atmosphere boundary layer (MABL) is the largest source of the water vapor (Trenberth et al. 2007) to form cloud and precipitation not only locally (vertically) but also to be transported laterally to the convectively active area as ITCZ (e.g. Sherwood et al. 2010), coastal area (e.g. Ogino et al. 2017), or convective envelopes such as typhoon, Bau-front, etc. The moisture above the MABL is also important regarding the convective activity in various scales (e.g. Mapes et al. 2006).

Past studies demonstrated that the water vapor in the lower troposphere varies spatiotemporally associated with the meso-scale convective features (e.g. Zipser 1977; Tompkins 2001). However the observational evidences of mesoscale variation of the water vapor, especially over the ocean, were limited in the past. lidar is capable to detect effective Raman signals only during nighttime which had been durable for continuous long-term observation on board Mirai for years. To the system, we added capacities to receive Raman-scattering signals from water vapor and nitrogen. The present study reports the early results from the system to retrieve vertical profile of the water vapor in the marine lower troposphere to resolve meso-scale variations which is difficult and/or costly by the classic radiosonde observations.

2. Observations and methods

2.1 Mirai Lidar System

The compact lidar system on board Mirai (hereafter Mirai lidar) is utilized. The system is installed inside the container on the top deck of the vessel (approximately 20 m above sea surface), as in Fig. 1, and points zenith direction to obtain the vertical profile.

The lidar system was designed to follow the Mie-scattering lidar system previously on board Mirai (Sugimoto et al. 2001) by adding Raman-observation capacity. The system transmits laser pulse at 1064 nm, 532 nm and 387 nm for nitrogen. The raw data are retrieved in bins sized 1 minute for temporal and 7.5 meters (Raman signals) or 6 meters (Mie signals) for vertical. Further details can be found in the Supplement. It should be noted that the Mirai lidar is capable to detect effective Raman signals only during nighttime without insolation.

2.2 Retrieval method

The Raman-backscattered signal strength can be assumed to be proportional to the number density of the molecular (Melfi et al. 1969). By assuming that the ratio of the nitrogen in the atmosphere is constant, we retrieve the water vapor mixing ratio from the ratio of the signal strength of the Raman scattering by water vapor to that by nitrogen.

We set the retrieval equation of lidar-based water vapor mixing ratio q(r) at range distance r as

\[ q(r) = \frac{K(r) Q_{w}(r) S_{w}(r) T_{w}(r)}{Q_{N} S_{N} T_{N}} \]  

where K(r) is the coefficient to convert the observed and corrected...
Finally determined as \( r \)'s. The median of the range distance. Obtained pairs of at 408 and 387 nm are utilized for products based on 355 nm laser signal (i.e. radiosonde-observed water vapor mixing ratio \( q_{rs} \)) to calculate the wavelength of Raman backscatter signal for molecular vapor mixing ratio, \( O_{wv} \) (Sakai et al. 2019). In the present study, \( q_{rs} \) is the transparency between sensor and range distance \( r \) at the wavelength of Raman backscatter signal for molecular \( x \).

The geometric occupancy of the obtained \( q_i \) products by taking moving median of for the \( (\text{median of } r) \) to convert all available instantaneous (in bins sized 1-minutes and 30-m) \( r \) to obtain \( q_i \). Some more details of each step are as follows.

To obtain \( S_r(r) \) at four Raman channels, the photon counting data are utilized¹, with the non-paralyzable photon pileup correction (Whiteman et al. 1992). The data are adopted when \( S_r(r) \) is strong enough and without contamination by cloud and precipitation. To reject weak signal, we set two thresholds. One is an absolute value to reject photon counting less than 10. Another is a relative value at each instantaneous profile, to reject the data bin where the photon counting is less than that beyond 7.5 km in range distance to regard as the background noises (lighting on the ship, moonlight, etc.). The presence of cloud and/or precipitation is determined by the Mic scattering signal at 1064 nm. Both Mic and Raman signals are integrated in 30-m range bins to match the vertical resolution. The received signals at 660 nm and 607 nm are used as \( S_x(r) \) and \( S_r(r) \), respectively, to calculate products based on scattering of laser at 532 nm (hereafter \( q_{23} \)). Likewise, signals at 408 and 387 nm are utilized for products based on 355 nm laser \( q_{355} \).

To calculate \( T_r(r) \) in the present study, we consider effects of the dry atmospheric gases, while effects of the aerosol is disregarded for simplicity. To obtain the gas attenuations, vertical profile of air temperature and pressure are obtained by interpolating COSPER International Reference Atmosphere (CIRA86; Fleming et al. 1988) at instantaneous ship location and date-in-year. \( O_{wv}(r) \) and \( O_{wv}(r) \) are assumed as identical, i.e. \( O_{wv}(r) = 1 \), to avoid complexity. However, the actual \( O \) can be affected by the different beam overlap, different optical path inside the observing system, and/or spatial inhomogeneity of the detector sensitivity (e.g. Sakai et al. 2019). In the present study, \( K(r) \) implicitly include these effects.

To determine \( K(r) \) by comparing \( SR \) to the collocating \( q_{23} \), we prepare special set of \( SR \) by taking moving median of \( SR \) for the window with the 10 minutes for the time and 120 m for the range distance, to suppress noise and the spatial sampling difference between lidar and radiosonde. \( q_i \) are also averaged for corresponding range distance. Obtained pairs of \( q_i \) and collocated \( SR \) lead to a set of \( K(r) \)'s. The median of the \( K(r) \)'s at each range distance is finally determined as \( K(r) \) to convert all available instantaneous (in bins sized 1-minutes and 30-m) \( r \) to obtain \( q_i \).

The geometric occupancy of the obtained \( q_i \), which originally aligns along the lidar ray path, are then converted to the height above sea level. The nadir angle of the ray path is provided by the shipboard gyro. The spatiotemporal resampling is also applied, by taking the median within the certain sampling window (ranges for time and height), when the number of valid data bin are greater than 2/3 of the number of all data bin. In the present study, we set the window as 10 minutes in temporal, and 120 m in vertical, to sufficiently resolve the meso-scale variations both in spatiotemporal scale and accuracy.

### 2.3 Cases and concurrent observations

Among the available observation periods, the present study focuses on the two cruises in the tropical eastern Indian Ocean for the project "Years of the Maritime Continent" (YMC; Yoneyama and Zhang 2020). The cruises are named as “MR15-04” and “MR17-08”, which are part of the field campaigns “Pre-YMC” and “YMC-Sumatra”, respectively. Yokoi et al. (2017, 2019) are referred to know the outline of the observation periods. The basic information of the data used in this study are summarized in Table 1.

During both periods, the radiosondes were launched every 3 hours. Furthermore, these two cruises include the periods of “stationary observation” when Mirai stayed within 4 kilometers from a reference point. These enable us to utilize larger number with smaller spatial difference from lidar. We included all available radiosonde profiles during the periods in Table 1 to better estimate \( K(r) \). On the other hand, the types of radiosonde sensor are different between the two periods (see Table 1). The observed relative humidity by these two types of sensors differ by 2% (Kawai et al. 2017). The nominal accuracies of relative humidity also differ by 1%. Furthermore the configurations of the Mirai lidar differ between two cruises (see Supplement). Considering these differences on both lidar and radiosonde, we determine individual \( K(r) \) for MR15-04 and MR17-08, and for \( q_{23} \) and \( q_{355} \) is available only in MR17-08.

The data from the surface meteorological package named “SOAR” and C-band radar on board Mirai are utilized in the case study (Section 4). The details of these data can be found in the “cruise report” for each cruise (JAMSTEC and BPPT, 2016, 2018).

### 3. Quantitative evaluation

This section examines \( q_i \) by comparing to \( q_{23} \). Though \( K(r) \) in Eq. (1) is obtained by referring \( q_{23} \), the individual \( q_i \) at each time and height is not guaranteed to match to \( q_{23} \).

The scattering diagrams to compare all available pair of \( q_i \) and corresponding \( q_i \) are shown in Fig. 2. For all three datasets, the correlation coefficients are high enough as expected. The root mean squared difference (RMSD) ranges 1.0–1.4 g/kg, which corresponds to approximately 5–10% of \( q_i \). These values are larger but comparable to those by the recent land-based mobile Raman lidar (Sakai et al. 2019) despite \( q_i \) is 1.5 times higher (i.e. higher optical thickness) for the present case.

Comparisons of \( q_i \) and \( q_{23} \) are extended to examine vertical profiles, as in Fig. 3. The averaged profiles (left panels in Fig. 3) well matches as expected, except around the highest part where the number of paired data are small (as in right panel in Fig. 3). The biases are generally smaller than 0.3 g/kg as in the panels at middle row of Fig. 3. Same panels indicate that the RMSD and quartiles of \( q_i \) and \( q_{23} \) are generally around or less than 1.0 g/kg in the

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1 A mechanical restriction did not allow us to apply the gluing method of photon counting and analog data (Whiteman et al. 2006) which is well accepted to be better quality.

| Name of Cruise | MR15-04 | MR17-08 |
|----------------|---------|---------|
| Period for the analyses in the present study | 23 Nov.–17 Dec. 2015 (25 days) | 30 Nov.–13 Dec. 2017 (14 days)¹ |
| Lidar data | Available \( q_{23} \) | \( q_{23} \) |
| Radiosonde data | Number of profiles when \( q_i \) is available | 92 | 56 |
| | Type of sensor | RS92-SGP, Vaisala | RS41, Vaisala |
| | Nominal accuracy of relative humidity² | 5% | 4% |

¹ Data after 13 December 2017 are omitted from the analyses in the present study due to insufficient quality of raw data from the Mirai lidar.

² According to the manufacturer, as in Kawai et al. (2017).
MABL (which can be found as less vertical gradient of averaged $q_l$ and $q_{rs}$ below 0.4 km height). The value is larger but comparable to the observed heterogeneity of moisture in the MABL (> 0.5 g/kg in Fabry 2006; ∼0.5 g/kg in Shinoda et al. 2009). These suggest that the large part of $q_l - q_{rs}$ possibly caused by the spatiotemporal sampling difference between lidar (spatiotemporal average of zenith-looking data) and radiosonde (instantaneous data along track of drifting balloon). On the other hand, the RMSD and interquartile range are larger with height. However the RMSD below 1.0 km height are generally ≤ 1.5 g/kg, which corresponds ≤ 10% of $q_{rs}$. These results indicate that the retrieval method in this study performs reasonably well and comparable to the land-based observations (e.g. Sakai et al. 2019) at least up to 1.0 km height which well covers the MABL.

Finally we compare $q_l_{532}$ and $q_l_{355}$. As in Fig. 2, both correlation coefficient and RMSD for all available data are close. However, Fig. 3 clearly shows that $q_l_{355}$ is retrieved to the higher altitude than $q_l_{532}$, without severe increase of the difference between $q_l - q_{rs}$. These indicates that, in the present case, $q_l_{355}$ can be better in performance than $q_l_{532}$, especially for the available height range.

4. Case study

A case study is performed to explore practicality of $q_l$ in focus on spatiotemporal variation in and above the MABL. The case is on 12 December 2017 when the westerly wind with westward propagating precipitating systems prevailed (Yokoi et al. 2019). The time-height cross section of $q_l_{532}$ and $q_l_{355}$ are shown in Fig. 4a and 4b, respectively. Both plots commonly show quick and large variations at around 14 UTC and 21 UTC, while $q_l_{355}$ is retrieved up to higher altitude than $q_l_{532}$ as shown in Fig. 3.

At 14 UTC, the $q_l$ in the MABL drops suddenly. The data missing between 1330 UTC to 1430 UTC, down to 0.3 km in height, indicates the low clouds existence, while the surface instruments did not detect any precipitation larger than its sensitivity (0.1 mm/h) (Fig. 4d). Therefore the non-precipitating low-clouds presumably passed over Mirai when $q_l$ in the MABL drops. The surface observation (Figs. 4c, 4d, and 4e) indicated that the time was at the end of 4-hour-long period (from past 10 UTC to past 14 UTC) with dropped temperature (down to 26.1°C at 1040 UTC in minimum) and gradually increasing water vapor mixing ratio ($q_{sfc}$) (up to 20.5 g/kg at 1350 UTC in maximum). The wind in 4-hour-long period was weak west-northwesterly while slightly-strong northwesterly before and after the period. Considering with those variations of the surface meteorological parameters, the sudden drop of $q_l$ in the MABL at 14 UTC was estimated as an edge of a meso-scale patch of cold and moist MABL, which accompanied non-precipitating clouds at its edge.

Before 21 UTC, the MABL vanished for a while, with the very dry air. It happened after the period with large missing of $q_l$ between 1940 to 2030 UTC. In the missing periods, Mie-scattering data (Fig. 4f) well captured gradually-rising intense signal, classified as the cloud base, with weak vertically-elongated signal around 2030 UTC, estimated as precipitation. The precipitation was also detected by the surface instruments (Fig. 4d). The surface observations also detected temperature drop around 1940 UTC (Fig. 4c) with increasing wind speed (especially in zonal compo-
Fig. 3. Comparison of the vertical profiles of water vapor mixing ratio. Panels in the left column ((a), (d) and (g)) are for the averaged profile of $q_l$ (red) and $q_r$, at corresponding time (black). Panels in the middle column are for the $q_l - q_r$, as in the average (thick black), average ± standard deviation (gray), median (thick red), and 1st and 3rd quartiles (thin red), respectively. Panels in the right column are number of available pairs of $q_l$ and $q_r$. Top panels are for $q_{532}$ in MR15-04, middle panels are for $q_{532}$ in MR17-08, and bottom panels are for $q_{355}$ in MR17-08.
mixing ratio for the lower troposphere over the ocean. The signal ratio of Raman water vapor to Raman nitrogen from the lidar are utilized to estimate water vapor mixing ratio ($q_l$) by referring the radiosonde-observed water vapor mixing ratio ($q_s$). Our demonstration involves data from two special observations, “MR15-04” cruise during “Pre-YMC” and “MR17-08” cruise during “YMC-Sumatra”. The obtained $q_l$ are shown to be quantitatively reasonable up to 1 km height that covers the MABL, and to be feasible to study meso-scale features of MABL and free troposphere above. $q_l$ from 355 nm laser (received signal at 387 and 408 nm) is advantageous to analyze up to higher altitude than that from 532 nm laser (received signal at 607 and 660 nm), while the quality resembles.

The results in the present study are the first step to observe detailed water vapor distribution in and above the MABL. Many possible future works are remained, nevertheless. The present study utilizes the best available radiosonde data (both in quality and frequency) among the various cruises to estimate the conversion coefficient from lidar signals to water vapor. The $q_l$ retrieval method with lesser number of, or without, radiosonde observation is essential to take advantage of lidar observation which provides continuous data with less human resources and expendables. The present study demonstrated the capability for tropical very wet and warm atmosphere, while the drier and/or colder conditions should be also examined to evaluate the dynamic range. The extension of the data in altitude (to higher altitude), in time (especially in daytime), and in quality, will contribute to progress further process studies in fine scale for, such as, vertical transport of the water vapor from the MABL to the free troposphere (e.g. to extend works by Bellenger et al. 2015), variations of sea surface latent heat flux associated to the moisture variation in MABL (e.g. to extend works by Chuda et al. 2008; Yokoi et al. 2014), etc. The data are also expected to validate products from present and future satellites including EarthCARE (Illingworth et al. 2015). The improvement both in software and hardware are desired in future works.

Acknowledgments

The authors would like to grateful to Drs. Ichiro Matsui and Atsushi Shimizu for their invaluable supports to establish both hardware and software of the Mirai lidar. The authors also thank who involved in field campaigns, especially crew and technical staff on board R/V Mirai, for their supports to obtain dataset. Comments and suggestions from the editor and two anonymous reviewers were essential to improve the manuscript. The study was supported in part by JAXA Research Announcement on the Earth Observations, and by JSPS KAKENHI Grant 17H04477.

Edited by: M. Yamanaka

Supplement

Outline of the Mirai lidar system during YMC field campaigns.

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Manuscript received 23 February 2020, accepted 1 May 2020 SOLA: https://www.jstage.jst.go.jp/browse/sola