

Observations of large aerosol infrared forcing at the surface

Andrew M. Vogelmann,¹ Piotr J. Flatau,^{1,2} Malgorzata Szczodrak,³
Krzysztof M. Markowicz,⁴ and Peter J. Minnett³

Received 22 December 2002; revised 12 May 2003; accepted 21 May 2003; published 28 June 2003.

[1] Studies of aerosol effects on the Earth's energy budget usually consider only the cooling effects at short (solar) wavelengths, but we demonstrate that they also have important warming effects at thermal infrared (IR) wavelengths that have rarely been observed and are commonly ignored in climate models. We use high-resolution spectra to obtain the IR radiative forcing at the surface for aerosols encountered in the outflow from northeastern Asia. The spectra were measured by the Marine-Atmospheric Emitted Radiance Interferometer (M-AERI) from the NOAA Ship *Ronald H. Brown* during the Aerosol Characterization Experiment-Asia (ACE-Asia). We show that the daytime surface IR forcing are often a few Wm^{-2} and can reach almost 10 Wm^{-2} for large aerosol loadings. Thus, even the smaller aerosol IR forcing observed here are comparable to or greater than the 1 to 2 Wm^{-2} IR surface enhancement from increases in greenhouse gases. These results highlight the importance of aerosol IR forcing which should be included in climate model simulations. **INDEX TERMS:** 0305 Atmospheric Composition and Structure: Aerosols and particles (0345, 4801); 0394 Atmospheric Composition and Structure: Instruments and techniques; 1630 Global Change: Impact phenomena; 3359 Meteorology and Atmospheric Dynamics: Radiative processes; 3360 Meteorology and Atmospheric Dynamics: Remote sensing. **Citation:** Vogelmann, A. M., P. J. Flatau, M. Szczodrak, K. M. Markowicz, and P. J. Minnett, Observations of large aerosol infrared forcing at the surface, *Geophys. Res. Lett.*, 30(12), 1655, doi:10.1029/2002GL016829, 2003.

1. Introduction

[2] The impact of aerosols on the Earth's energy budget can be quantified in terms of aerosol radiative forcing, defined as the difference between atmospheric radiative fluxes when aerosols are present and when they are absent. Compared to the aerosol forcing at short wavelengths, the forcing at IR wavelengths has received much less consideration, and modeling aerosol IR effects contains many uncertainties [Andreae, 1996; Sokolik and Toon, 1996; Tegen et al., 1996]. For example, mineral aerosols have

complex, highly varied optical properties that, for equal loadings, can cause differences in the surface IR flux between 7 and 25 Wm^{-2} [Sokolik et al., 1998]. Consequently, only a few large-scale climate models currently consider aerosol IR effects [e.g., Tegen et al., 1996; Jacobson, 2001] despite their potentially large forcing. For instance, calculations by Lubin et al. [2002] show that the aerosol IR radiative forcing could be as much as 15% of the large, anthropogenic shortwave forcing found during the Indian Ocean Experiment (INDOEX). Direct observations of aerosol IR forcing are rare at either the surface [e.g., Lubin and Simpson, 1994; Spänkuch et al., 2000], in the atmosphere [Highwood et al., 2003], or at the top of the atmosphere [Ackerman and Chung, 1992; Hsu et al., 2000]. Thus, our understanding of IR aerosol forcing is largely model-based (with the unknowns therein) with few direct observations available for validation. We address this issue using a survey of data obtained from the NOAA Ship *Ronald H. Brown* during the Aerosol Characterization Experiment-Asia (ACE-Asia) conducted in the spring of 2001 (B. J. Huebert et al., An overview of ACE-Asia: strategies for quantifying the relationships between Asian aerosols and their climatic impacts, submitted to the *Journal of Geophysical Research*, 2003).

2. Cruise Data

[3] The NOAA Ship *Ronald H. Brown* cruise departed from Honolulu, Hawaii on 15 March, 2001, traversed the Pacific to enter the Sea of Japan (East Sea) via the Korea Strait, and returned to end at Yokosuka, Japan on 20 April. The ship encountered a range of aerosol types and column amounts as the ship travelled from the clean ocean air near Hawaii to the aerosol-laden skies over the Sea of Japan. Comprehensive measurements were made of the downwelling atmospheric radiation, and of the aerosol properties. The shortwave component of the aerosol forcing were analysed using these data in other studies [e.g., Markowicz et al., 2003a].

[4] The Marine-Atmospheric Emitted Radiance Interferometer (M-AERI) is a shipboard, Fourier Transform Infrared (FTIR) radiometer [Minnett et al., 2001]. It measured IR sky radiance spectra from 520 to 3020 cm^{-1} (about 18 to $3 \mu\text{m}$) at $\sim 0.5 \text{ cm}^{-1}$ resolution, with a narrow field-of-view (45 mrad , full angle) at 0° (overhead) and 55° to zenith. Each cloud-free spectrum contains the combined downward emissions by aerosols and greenhouse gases (e.g., H_2O , CO_2). Direct observational determinations of aerosol IR forcing are scarce because of the difficulty in isolating the aerosol signal from the background gaseous emissions. Here we make use of several M-AERI features that make this determination possible: 1) it has two on-board black body references, which produce a real-time calibration that

¹Center for Atmospheric Sciences, Scripps Institution of Oceanography, University of California at San Diego, La Jolla, California, USA.

²UCAR Visiting Scientist to the Naval Research Laboratory, Monterey, California, USA.

³Rosenstiel School of Marine and Atmospheric Science, Division of Meteorology and Physical Oceanography, University of Miami, Miami, Florida, USA.

⁴Institute of Geophysics, University of Warsaw, Warsaw, Poland.

is accurate to better than 1% of the ambient radiance; 2) it has a fast measurement frequency, providing a spectrum every 10 min; and 3) its high spectral resolution enables isolating the atmospheric windows, where gaseous emission is weak enough for aerosol IR forcing to be significant (outside the windows the aerosol emission is swamped by strong gaseous emissions [Ackerman *et al.*, 1976]). In addition, the M-AERI radiance spectra can be inverted to retrieve the water vapor and temperature profiles for the lowest 3 km of the atmosphere [Smith *et al.*, 1999; Feltz *et al.*, 1998]. These retrievals are important for separating the IR signatures of the lower atmospheric aerosols from the variations in the atmospheric background caused by fluctuations in the temperature and water vapor profiles. The iterative retrieval process requires an initial guess of the temperature and moisture profiles, which is obtained from radiosondes that were launched from the ship two to five times a day.

3. Methodology

[5] The aerosol's radiative effect is obtained from a M-AERI spectrum by removing the background gaseous emission. This is achieved by subtracting a clean-sky (no aerosol) spectrum that is modeled using concurrent observations of the gaseous state [e.g., Lubin and Simpson, 1994; Spänkuch *et al.*, 2000; Highwood *et al.*, 2003], i.e.,

$$\text{Aerosol Radiance Effect} = (M - \text{AERI Observed Radiance}) \\ - (\text{Modeled Clean - Sky Radiance})$$

This aerosol radiance effect is converted to an aerosol IR radiative forcing (flux) using a radiance-to-flux conversion. We focus on the 10- μm atmospheric window, where most (about 87%) of the aerosol forcing occurs. Note that the aerosol IR forcing thus obtained is a bulk, energetic quantity that implicitly includes the net surface radiative effects of aerosol amount, composition, size, height and temperature. This is different from aerosol retrievals, where the effects of the intermediate atmosphere would be removed. An advantage of using IR forcing is that it directly conveys the net IR effect of aerosol as seen in the surface energy balance.

[6] The clean-sky (aerosol-free) spectra of the gaseous emissions are computed at high resolution using the Line-By-Line Radiative Transfer Model [LBLRTM; Clough and Iacono, 1995]. LBLRTM has been validated and improved using AERI data [e.g., Tobin *et al.*, 1999]. We use a recent release of LBLRTM (v6.01 with v2.4.1 of the CKD water vapor continuum) with the HITRAN database [Rothman *et al.*, 1998] that provides the spectroscopic line parameters (HITRAN-2000 v11, updated through 4/2001). Our line-by-line model spectra are made comparable to the observed spectra by applying a sinc scanning function using the M-AERI maximum optical path difference (length of an equivalent interferometer) of $\pm 1.037 \text{ cm}^{-1}$ [Minnett *et al.*, 2001].

[7] The model requires atmospheric profiles of temperature, and of the concentrations of water vapor and ozone. The model profiles of water vapor and temperature are specified every 10 min using M-AERI retrievals for the

lowest 3 km and, above, by the nearest radiosonde launched from the ship. This representation captures the high-frequency variability of the lower atmosphere, which contains the bulk of the water vapor column and aerosol amount. (Note that the wavelengths used by the retrieval are in strong emission regions that are outside the window region used for the aerosol studies.) The ozone model inputs are syntheses of: surface ozone concentrations, measured from the Ronald H. Brown every 30 min; ozonesondes, launched from nearby islands; and the total ozone column, measured by a Microtops Ozonometer [Morys *et al.*, 2001]. The boundary layer profile of the nearest ozonesonde is scaled to match the concentration of the lowest level to the surface ozone measurement, and the upper atmospheric ozone concentrations are scaled to match the total column measured by the Microtops. Uncertainties in the ozone profile can lead to conspicuous spectral features (e.g., a sharp 'trough' or 'finger') appearing in the radiance residuals of the 1000–1050 cm^{-1} ozone band. The radiative flux within such a feature is small, usually $<0.5 \text{ Wm}^{-2}$, and the ozone profile is tuned to remove it.

[8] The aerosol radiance effects are obtained for each view angle (0° and 55°) by subtracting from the M-AERI spectra the modeled, clean-sky spectra and spectrally integrating each from 700 to 1350 cm^{-1} , which encompasses the 10- μm window region. A regression equation provides the angular integration that translates these two radiance quantities into the aerosol IR radiative forcing (flux). The regression coefficients are computed using the MODTRAN-v4 multiple-scattering radiative transfer algorithm [Anderson *et al.*, 2001]. When the two view angles are used, the regression coefficients are independent of the aerosol type and amount, and of the boundary layer water vapor amount. The flux conversions are accurate to within 0.1 Wm^{-2} for most cases and 0.25 Wm^{-2} for all cases.

[9] Cloud-free periods are determined for the M-AERI narrow field-of-view by examinations of the lidar vertical profiles, Total Sky Imager data, and radiosonde profiles. The candidate cases are screened further to ensure that thin clouds or other artifacts do not contaminate the analysis. A cloud has a spectral shape and magnitude that are distinctively different from the aerosols (not shown), and the case is rejected if its radiance effect contains a cloud spectral signature. It is also rejected if the integrated radiances at 0° are significantly larger than 55° , or if gross inconsistencies are found between the spectral features at 0° and 55° . Both occurrences are physically inconsistent with the atmosphere being effectively plane-parallel, which is required for using the radiance-to-flux regression equation. Finally, the accuracies of the M-AERI water vapor retrievals are considered. Consistent with a comparison between water vapor profiles determined by AERI retrievals and by a Raman lidar [Turner *et al.*, 2000], we find that the sondes and M-AERI retrievals usually agree to within 2% to -5% of the boundary layer water vapor column. However, larger deviations are accompanied by a 'spike' in the radiance residual near 800 cm^{-1} . The occurrence of this spectral feature in the analysis suggests that the M-AERI retrieval is suspect and the case is rejected.

[10] The uncertainty of the deduced aerosol IR forcing is quantified by examining the forcing for days with low

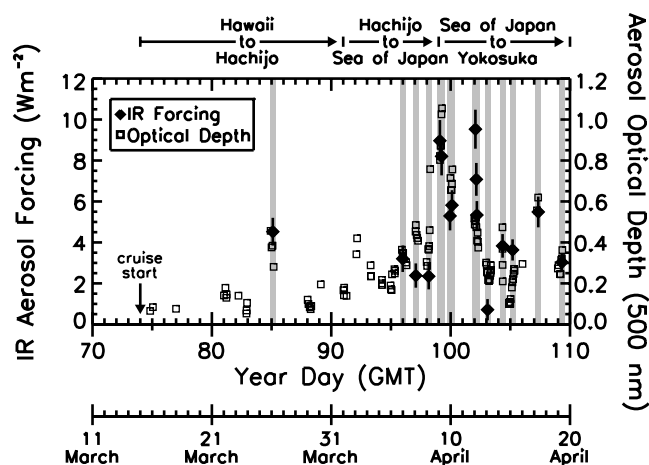


Figure 1. Observationally determined aerosol IR forcing for the cruise. Left axis gives the (daytime) aerosol IR surface forcing ($\blacklozenge$); right axis gives the aerosol optical depth at 500 nm ($\square$). Shading indicates days for which the aerosol IR forcing is evaluated, which is where optical depths are greater than about 0.3 and a converged M-AERI retrieval exists. Error bars for the aerosol IR forcing are based on the analysis described in the text. Uncertainties in the optical depths are about 0.01 (not plotted). Cruise transit locations given above the plot. Hachijo Island is off the southeast coast of Japan; the Port of Yokosuka is located at the entrance to Tokyo Bay. The *Ronald H. Brown* air mass classification indicates signs of continental aerosols at the surface beginning YD 84.

aerosol depths. The values for such days should be approximately zero; so a nonzero ‘trace’ provides an uncertainty estimate of the analysis procedure. We find a small trace that trends slightly with water vapor content, and we ‘calibrate’ the deduced IR aerosol forcing by subtracting this trace. (Note that this trace implicitly includes the uncertainty in the M-AERI water vapor retrievals.) Thus, the total uncertainty of the aerosol IR forcing is the combined uncertainties of the trace, the ozone correction, and the radiance-to-flux conversion.

4. Results

[11] The aerosol amount and IR forcing varied considerably during the cruise (Figure 1). The column aerosol amounts are quantified here in terms of the aerosol optical depth at 500 nm (mid-visible). The optical depths were measured for cloud-free periods using a hand-held Microtops Sun photometer [Morys *et al.*, 2001], which was calibrated via Langley regressions performed at the Mauna Loa Observatory. The optical depths increased dramatically as the ship approached Japan and, particularly, when it entered the Sea of Japan (Figure 1). Many of the aerosol optical depths are greater than 0.4, and a few even reach about 1.0.

[12] The associated daytime aerosol IR forcing shown can be large (Figure 1), with values ranging from a few Wm^{-2} to a maximum of $9.5 \pm 1.0 \text{ Wm}^{-2}$. The larger forcing are comparable to or greater than the 7.7 Wm^{-2} observed for the Los Angeles pollution [Lubin and Simp-

son, 1994]. Because infrared fluxes do not have the same kind of pronounced diurnal cycle as shortwave fluxes, these instantaneous IR aerosol forcing serve as estimates of the longer-term average of the IR forcing. Our analysis is restricted to daytime values because we use visible data for cloud screening. Nighttime values could be different because of thermal effects resulting from aerosol solar heating during the daytime, and nighttime relative humidity effects on aerosol size that might enhance the aerosol emission, contingent on factors such as aerosol advection and height changes.

[13] Figure 1 indicates that the magnitude of the IR forcing relative to the visible optical depth (i.e., their ratio) varies throughout this period. For example, the IR forcing for Year Day (YD) 100 and at the beginning of YD 102 are similar, but the visible optical depth for YD 100 is double that for YD 102. This suggests that, for these days, a dramatic difference exists in the aerosol infrared optical depths that are not revealed in the visible optical depths. Such differences are manifestations of the differences in optical properties with aerosol type, and particle size. For example, dust is known to be particularly efficient (per visible optical depth) at producing large surface IR forcing [Tegen *et al.*, 1996; Sokolik *et al.*, 1998], and the boundary layer for YD 102 contained large concentrations of dust that changed throughout the day. Thus, accurately modeling the aerosol IR forcing requires an adequate specification of the variations of these aerosol properties within the atmospheric column. Details of how factors such as aerosol height, size, and water vapor amount affect the IR forcing are diagnosed in a separate modeling study [Markowicz *et al.*, 2003b]. Further, within that study, these observed IR forcing have been modeled with good agreement using independent measurements of the aerosol chemical and optical properties made during the cruise. This supports the veracity of the methodology used to extract the aerosol IR forcing from the observed spectra.

5. Discussion

[14] This view of the surface component of the regional energy balance provides an essential complement to that offered by top-of-atmosphere (TOA) satellite studies [e.g., Ackerman and Chung, 1992; Hsu *et al.*, 2000] because the climate response depends not only on the energy balance at the TOA, but also on the heating at the surface and its atmospheric profile. The magnitude of these daytime observed IR surface forcing are significant. The largest IR forcings represent an increase of about 20 to 40% (depending on the column water vapor amount) of the aerosol-free flux in the central portion of the window ($800\text{--}1250 \text{ cm}^{-1}$). Furthermore, the forcing observed here are comparable to or larger than the 1 to 2 Wm^{-2} change in the globally averaged surface IR forcing caused by greenhouse gas increases since pre-industrial times [Tegen *et al.*, 1996; Ramanathan *et al.*, 2001]. (This surface greenhouse enhancement should not be confused with the value at the tropopause that is commonly used in climate studies.) Of course the episodic nature of the aerosol forcing do not have the constancy of greenhouse gas forcing but, during this period, such episodes were prevalent within northeastern Asia. Thus, aerosol IR forcing of the magnitudes observed

here modify the surface energy balance by a magnitude that may be important for the climate. Note that by comparing the surface forcing by aerosol to that by greenhouse gases, we do not intend to imply that they share the same degree of anthropogenic origin. Only that, without including aerosol IR forcing, models used for climate and climate change studies may be missing an important forcing, regardless of whether the aerosol source is natural or anthropogenic.

[15] In summary, much attention has been devoted to the solar forcing by aerosols, but the observational results presented here underscore the importance of aerosol IR forcing and resolving uncertainties therein [Andreae, 1996; Sokolik and Toon, 1996; Sokolik et al., 1998; Miller and Tegen, 1999]. Unravelling climate interactions, especially over for long-term climate trends, requires ascertaining and treating all of the radiatively important components of the climate system. Consistent with prior modeling studies [e.g., Tegen et al., 1996; Sokolik et al., 1998] our observational analysis indicates that aerosol IR forcing can be significant and, thus, should be included in climate model simulations.

[16] **Acknowledgments.** We thank members of the ACE-Asia science team for their coordination and planning. K. Maillet, U. Miami, is thanked for his at-sea support. We acknowledge helpful discussions with V. Ramanathan, D. Lubin, R. von Glasow, S. A. Clough, and with W. Feltz, who also provided a timely update of the retrieval database. We thank two reviewers for their constructive comments. We are grateful for the surface ozone measurements provided by J. Johnson, and for the ozonesondes provided by S. Oltmans, J. Kim and, for TRACE-P, H. Sasaki for the Japanese Meteorological Agency. PMEL provided the air mass classifications and back trajectories used. Work by A. M. Vogelmann was supported by NSF (ATM-0109135) and by the Office of Biological and Environmental Research of the U.S. Department of Energy (DE-FG03-03ER63538) as part of the Atmospheric Radiation Measurement Program. P. J. Flatau was supported by the NSF Climate Dynamics Program and ONR Program Element PE0602435N, P. J. Minnett and M. Szczodrak by NASA (NAS5-31361), and K. Markowicz was supported by a Fulbright Grant. This research is a contribution to the International Global Atmospheric Chemistry (IGAC) Core Project of the International Geosphere Biosphere Program (IGBP) and is part of the IGAC Aerosol Characterization Experiments (ACE).

References

- Ackerman, S. A., and H. Chung, Radiative effects of airborne dust on regional energy budgets at the top of the atmosphere, *J. Appl. Meteor.*, **31**, 223–233, 1992.
- Ackerman, T. P., K.-N. Liou, and C. B. Leovy, Infrared radiative transfer in polluted atmospheres, *J. Appl. Meteor.*, **15**, 28–35, 1976.
- Anderson, G. P., A. Berk, P. K. Acharya, M. W. Matthew, L. S. Bernstein, J. H. Chetwynd, H. Dothe, S. M. Adler-Golden, A. J. Ratkowski, G. W. Felde, J. A. Gardner, M. L. Hoke, S. C. Richtsmeier, and L. S. Jeong, MODTRAN4, version 2: Radiative transfer modeling, *SPIE-Int. Soc. Opt. Eng.*, **4381**, 455–459, 2001.
- Andreae, M. O., Raising dust in the greenhouse, *Nature*, **380**, 389–390, 1996.
- Clough, S. A., and M. J. Iacono, Line-by-line calculation of atmospheric fluxes and cooling rates. 2. Application to carbon dioxide, ozone, methane, nitrous oxide and the halocarbons, *J. Geophys. Res.*, **100**, 16,519–16,535, 1995.
- Feltz, W. F., W. L. Smith Knuteson, R. O. Revercomb, H. M. Woolf, and H. B. Howell, Meteorological applications of temperature and water vapor retrievals from the ground-based Atmospheric Emitted Radiance Interferometer (AERI), *J. Appl. Meteor.*, **37**, 857–875, 1998.
- Hsu, N. C., J. R. Herman, and C. J. Weaver, Determination of radiative forcing of Saharan dust using combined TOMS and ERBE data, *J. Geophys. Res.*, **105**, 20,649–20,661, 2000.
- Highwood, E. J., J. M. Haywood, M. D. Silverstone, S. M. Newman, and J. P. Taylor, Radiative properties and direct effect of Saharan dust measured by the C-130 aircraft during SHADE.2: Terrestrial spectrum, *J. Geophys. Res.*, in press, 2003.
- Jacobson, M. Z., Global direct radiative forcing due to multicomponent anthropogenic and natural aerosols, *J. Geophys. Res.*, **106**, 1551–1568, 2001.
- Lubin, D., and A. S. Simpson, The longwave emission signature of urban pollution: Radiometric FTIR measurement, *Geophys. Res. Lett.*, **21**, 37–40, 1994.
- Lubin, D., S.-K. Satheesh, G. McFarquhar, and A. J. Heymsfield, Long-wave radiative forcing of Indian Ocean tropospheric aerosol, *J. Geophys. Res.*, **107**(D19), 10.1029/2001JD001183, 2002.
- Markowicz, K. M., P. J. Flatau, P. K. Quinn, C. M. Carrico, M. K. Flatau, A. M. Vogelmann, D. Bates, M. Liu, and M. J. Rood, Influence of relative humidity on aerosol radiative forcing: An ACE-Asia Experiment Perspective, *J. Geophys. Res.*, in press, 2003a.
- Markowicz, K. M., P. J. Flatau, A. M. Vogelmann, P. K. Quinn, and E. J. Welton, Clear-sky infrared aerosol radiative forcing at the surface and the top of the atmosphere, *Q. J. Roy. Met. Soc.*, in press, 2003b.
- Miller, R. L., and I. Tegen, Radiative forcing of a tropical direct circulation by soil dust aerosols, *J. Atmos. Sci.*, **56**, 2403–2433, 1999.
- Minnett, P. J., R. O. Knuteson, F. A. Best, B. J. Osborne, J. A. Hanafin, and O. B. Brown, The Maritime-Atmospheric Emitted Radiance Interferometer: A high-accuracy, seagoing infrared spectroradiometer, *J. Atmos. Oceanic Tech.*, **18**, 994–1013, 2001.
- Morys, M., F. M. Mims, S. Hagerup, S. E. Anderson, A. Baker, J. Kia, and T. Walkup, Design, calibration, and performance of MICROTIPS II handheld ozone monitor and Sun photometer, *J. Geophys. Res.*, **106**, 14,573–14,582, 2001.
- Ramanathan, V., P. J. Crutzen, J. T. Kiehl, and D. Rosenfeld, Aerosols, climate, and the hydrological cycle, *Science*, **294**, 2119–2124, 2001.
- Rothman, L. S., et al., The HITRAN molecular spectroscopic database and HAWKS (HITRAN atmospheric workstation): 1996 edition, *J. Quant. Spectro. Radiat. Trans.*, **60**, 665–710, 1998.
- Smith, W. L., W. F. Feltz, R. O. Knuteson, H. E. Revercomb, H. M. Woolf, and H. B. Howell, The retrieval of planetary boundary layer structure using ground-based infrared spectral radiance measurements, *J. Atmos. Oceanic Technol.*, **16**, 323–333, 1999.
- Sokolik, I. N., and O. B. Toon, Direct radiative forcing by anthropogenic airborne mineral aerosols, *Nature*, **381**, 681–683, 1996.
- Sokolik, I. N., O. B. Toon, and R. W. Bergstrom, Modeling the radiative characteristics of airborne mineral aerosols at infrared wavelengths, *J. Geophys. Res.*, **103**, 8813–8826, 1998.
- Spänkuch, D., W. Dohler, and J. Guldner, Effect of coarse biogenic aerosol on downwelling infrared flux at the surface, *J. Geophys. Res.*, **105**, 17,341–17,350, 2000.
- Tegen, I., A. A. Lacis, and I. Fung, The influence on climate forcing of mineral aerosols from disturbed soils, *Nature*, **380**, 419–422, 1996.
- Tobin, D. C., F. A. Best, P. D. Brown, S. A. Clough, R. G. Dedeker, R. G. Ellingson, R. K. Garcia, H. B. Howell, R. O. Knuteson, E. J. Mlawer, H. E. Revercomb, J. F. Short, P. F. W. van Delst, and V. P. Walden, Downwelling spectral radiance observations at the SHEBA ice station: Water vapor continuum measurements from 17 to 26 μm , *J. Geophys. Res.*, **104**, 2081–2092, 1999.
- Turner, D. D., W. F. Feltz, and R. A. Ferrare, Continuous Water Vapor Profiles from Operational Ground-Based Active and Passive Remote Sensors, *Bull. Amer. Meteor. Soc.*, **81**, 1301–1318, 2000.

A. M. Vogelmann and P. J. Flatau, Center for Atmospheric Sciences, Scripps Institution of Oceanography, University of California at San Diego, 9500 Gilman Drive, La Jolla, CA 92093-0221 USA. (avogelmann@ucsd.edu)

M. Szczodrak and P. J. Minnett, Rosenstiel School of Marine and Atmospheric Science, Division of Meteorology and Physical Oceanography, University of Miami, Miami, FL, USA.

K. M. Markowicz, Institute of Geophysics, University of Warsaw, Warsaw, Poland.