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The effects of anomalous atmospheres on the accuracy of infrared sea-surface temperature retrievals: Dry air layer intrusions over the tropical ocean

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ABSTRACT

The effects of layers in the atmosphere with anomalously low moisture content on the accuracy of the sea-surface temperature (SST) derived from measurements of infrared radiometers on earth observation satellites are quantified using measurements taken from research cruises and numerical simulations. Radiosonde data from areas of the oceans that are seasonally affected by intrusions of dry air masses of continental origin were used with the Line-By-Line Radiative Transfer Model (LBLRTM) to simulate brightness temperatures measured in MODIS bands 31 and 32 (at wavelengths of ~11 μm and ~12 μm). The radiosonde datasets contained profiles with and without a dry layer, representing the baseline (no dry layer aloft) and the anomalous conditions (dry layer present). MODIS SST retrieval algorithm versions 5 and 6 were applied to the simulated brightness temperatures to obtain the SST and the retrieval errors were examined. Whereas the average errors for the baseline 'no dry layer' conditions range between 0.29 and 0.72 K, in the case of very deep dry layers the errors can be >1 K. Simulations were also performed for atmospheric profiles that were created from the measured profiles by 'drying' layers at various altitudes. It was found that the retrieval errors 1) depend on the amount of water vapor in the atmosphere, 2) change in a systematic way dependent on the presence and characteristics of a dry layer, 3) dry layers in the lower troposphere make the SST retrieval errors more positive, and dry layers in the upper troposphere tend to introduce SST retrieval errors that are more negative.

1. Introduction

There are several sources of contributions to the uncertainties in sea-surface temperature (SST) retrievals from infrared radiometers on satellites, and one of the most important is the accuracy with which the effects of the intervening atmosphere are corrected. Other sources include the uncertainties in the radiometric measurements of the spacecraft instrument, and the effectiveness of the identification of measurements that include radiances from clouds and aerosols. The presence of clouds in the field of view of an imaging infrared radiometer is a major hindrance to the retrieval of SST and corrections cannot generally be made, which leads to the exclusion of the cloud-contaminated measurements from the SST retrieval process and can result in the rejection of ~80–90% of the infrared measurements (Kilpatrick, Podestá, & Evans, 2001). For aerosols, there is some evidence that corrections can be made (Merchant, Embury, Le Borgne, & Bellec, 2006; Nalli & Stowe, 2002), but as with clouds, it is usually better to attempt to identify their presence and discard the data from the SST retrieval process. The instrumental artifacts comprise those that are essentially irreducible, such as the noise generated by the detectors and imperfections in the digitizers, and those for which corrections can be derived given improved information about the characteristics of the instrument, often gained from housekeeping data and patterns in the errors in the derived variables (e.g. Xiong, Barnes, Guenther, & Murphy, 2003). Thus, the contribution to the uncertainties that is perhaps the most susceptible to improvement is that from the atmospheric correction.

In the cases where the atmosphere is free of cloud and aerosols, the radiative effects are due to the absorption and re-emission of infrared photons by molecules in the atmosphere (see 3.4 of Kidder and Haar (1995), or 4.5 of Martin (2004)). Since the atmosphere is nearly everywhere cooler than the sea surface, the consequence of the absorption and re-emission is to make the temperatures measured by the satellite radiometers, the brightness temperatures, cooler than the SST, and the purpose of the atmospheric correction algorithm is to compensate for this temperature deficit (e.g. Závody et al., 1995). Surface temperature retrievals from satellite radiometers are made in the wavelength intervals in the atmosphere's transmission spectrum where the transmissivity is high, the so-called atmospheric "windows", so that most of the signal originates at the surface and the temperature deficit is relatively small. In these spectral regions, such as in the wavelength range of ~10–12.5 μm, the main atmospheric constituent that interacts with the infrared radiation is water vapor; and water vapor is very variable in both space and time. Exploiting the wavelength dependence of the

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water vapor effect, an atmospheric correction algorithm can be devised based on the different brightness temperatures at different wavelengths in the atmospheric window (McMillin, 1975; Minnett & Barton, 2010, and references therein). Such atmospheric correction algorithms have sets of coefficients that are generally derived empirically, through the statistical analysis of "matchups" between the satellite brightness temperature measurements and collocated, contemporaneous measurements from surface thermometers or radiometers (e.g. Kilpatrick et al., 2001). An alternative method to derive the coefficients uses radiative transfer modeling to simulate the propagation of the spectrum of infrared emission from the sea-surface through a large number of atmospheric temperature and humidity profiles taken from radiosondes launch from ships or coastal stations, or from the representation of the marine atmosphere in the reanalysis fields of Numerical Weather Prediction Models. The relative spectral response functions of the satellite radiometer are used with the simulated spectra of infrared top-of-atmosphere emission to derive a large set of simulated brightness temperatures (Llewellyn-Jones, Minnett, Saunders, & Závody, 1984; Závody et al., 1995). The simulated top-of-atmosphere brightness temperatures and their associated SSTs are then subjected to the same statistical analysis as the "matchups" to derive coefficients for the atmospheric correction algorithm. In the empirical method there is an implicit assumption that the distribution of atmospheric properties in the "matchups" is a good representation of the range of cloud- and aerosol-free atmospheric conditions to which the algorithm will be applied. Similarly, in the modeling approach, there is the assumption that the selection of atmospheric temperature and humidity profiles used to generate the simulated brightness temperatures provides an appropriate representation of the atmospheric conditions for the retrievals.

In applying the atmospheric correction algorithms to generate global SST fields, it is to be expected that the uncertainties in the individual retrievals will have a distribution, probably close to Gaussian, that reflects the deviations of the atmospheric conditions from the mean of the distribution of those used to generate the coefficients used in the algorithm. The bigger the deviation, the bigger the SST retrieval error (Minnett, 1986). One feature that constitutes anomalous conditions, and therefore can lead to larger SST errors, is the presence of an atmospheric dry layer. Even if such layers exist in the selection of atmospheres used in the coefficient derivation, they tend to be regionally or seasonally constrained, and therefore will lead to uncertainties in the SSTs that have a regional and seasonal character.

Barton (2011) discusses a dependence of the SST retrieval error for a number of satellite SST sensors on the vertical distribution of water vapor. Barton expresses this by the partition of water vapor between lower and upper troposphere. In effect, dry layers are special (extreme) cases of the vertical distribution of water vapor in the atmosphere so one might suspect that under such conditions the SST error might be large.

A well-known and sometimes extreme dry layer is the Saharan Air Layer (SAL) that extends seasonally from the West African coast across the tropical North Atlantic Ocean (Prospero, 1999; Prospero & Carlson, 1972), often, but not necessarily always, associated with mineral aerosols (Zhang & Pennington, 2004). One consequence of the SAL is the suppression of the growth of Atlantic hurricanes (Dunion & Velden, 2004), and the identification of the SAL for use in hurricane forecasting is facilitated by the anomalous reduction in the satellite measurement of brightness temperature differences between measurements at 10.7 and 12 μm (Dunion & Velden, 2004), the same pair of spectral measurements used in the atmospheric correction for SST retrievals. Other occurrences of a dry layer of continental origin extending over the tropical ocean are off Northern Australia, and an example of this was seen in the Tropical Warm Pool International Cloud Experiment (TWP-ICE) held in the austral summer at the beginning of 2006 (May et al., 2008). An extensive cyclonic feature centered over the Northern Territory caused the winds to advect dry air at middle levels in the troposphere over the Timor Sea.

Elsewhere, throughout the tropics and subtropics, dry layers result as airmasses that have lost moisture through condensation events descend over intervals of days to lower levels, achieving low relative humidities, <20%, by adiabatic warming. These tend to form zonal bands in the tropics of both hemispheres (Cau, Methven, & Hoskins, 2007).

This study is directed to improving our understanding of the effects of dry layers on the uncertainties of SSTs derived from infrared radiometers on polar orbiting spacecraft, and is focused on events captured in ship-based measurements in the tropical Atlantic Ocean (SAL events) and in the Timor Sea (during the TWP-ICE campaign).

The algorithm used in retrieving SST values from top-of-atmosphere radiance measurements of infrared radiometers, such as the Advanced Very High Resolution Radiometer (Cracknell, 1997) and the MODerate-resolution Imaging Spectroradiometer (Esaias et al., 1998) on earth-observation satellites is based on statistical relationships between the brightness temperatures and the corresponding SST. A commonly used version is the Non-Linear SST algorithm (Walton, 1988) and was originally developed for AVHRR (May, Parmeter, Olszewski, & McKenzie, 1998). The MODIS version of the formula has the following form:

$$SST = a_1 + a_2T_{31} + a_3(T_{31} - T_{32})T_{ref} + a_4(T_{31} - T_{32})(\sec(\zeta) - 1) + \delta \quad (1)$$

where T_{31} and T_{32} are brightness temperatures (BT) in MODIS channels 31 and 32 (wavelength bandwidths of 10.78 to 11.28 μm and 11.77 to 12.27 μm), T_{ref} is a reference or "first guess" temperature, which is often taken as the "Reynolds" Optimally Interpolated SST (Reynolds et al., 2007), ζ is the satellite zenith angle, a_1 to a_4 are coefficients derived from coincident measurements of subsurface temperatures from buoys, and δ is a correction for the thermal skin effect so the retrieval is a skin SST. The coefficients are dependent on the relative spectral responses of the individual radiometers, through the conversion from measured spectral radiance to the equivalent top-of-atmosphere (TOA) BT, and on atmospheric variability, which results from the dependence of atmospheric absorption and emission on variable constituents of the atmosphere, especially water vapor. The second term on the right-hand side of Eq. (1) represents the first order effect relating the measured brightness temperature to that of the emitting sea surface. The third term represents the correction for the absorption and emission of the intervening atmosphere, primarily due to water vapor, and the fourth term accounts for the increased length of a slant path the radiation propagates through the atmosphere for off-nadir measurements. The skin effect correction δ accounts for the difference between the subsurface SST, as measured by the buoys, and the skin SST. The MODIS algorithms include the global average δ of −0.17 K (Donlon et al., 2002; Minnett, Smith, & Ward, 2011) so that the MODIS retrievals are estimates of the skin SST.

In the current version of the MODIS SST retrieval algorithm (Version 5, referred to here as V5) the '*a*' coefficients are derived separately for each month of the year from "matchups" between the MODIS BTs and subsurface temperature measurements from drifting and moored buoys. To avoid discontinuities at the start of each month a sliding temporal tapered weighting function over five months is used. Following experience gained from the AVHRR Pathfinder program (Kilpatrick et al., 2001), the coefficients are derived separately for two populations of the BT difference between MODIS channels 31 and 32 (ΔBT_{31-32}), which can be used a proxy for the precipitable water vapor (*PWV*). One set of coefficients is derived for $\Delta BT_{31-32} < 0.7$ K and other for $\Delta BT_{31-32} > 0.7$ K, corresponding to dry and moist atmospheres — this will be discussed further below. When the algorithm is applied to MODIS measurements, the "dry" coefficients are applied where $\Delta BT_{31-32} < 0.5$ K and "moist" for $\Delta BT_{31-32} > 0.9$ K; and a linear combination is used for the range between. The new Version 6 (V6) of the algorithm that has been recently developed has coefficients depending not only on the month of the year and ΔBT_{31-32}, but also on six latitude bands of the measurement to better capture the regional variability of atmospheric conditions between the sea level and the satellite (Evans, Minnett, & Podestá, 2013). The latitude zones are 0°–20°, 20°–40° and