Spatial variability of the atmosphere over southern England, and its effect on scene-based atmospheric corrections

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Earth observation data acquired in the optical region require atmospheric correction before they can be used quantitatively. Most operational methods of atmospheric correction assume that the atmospheric properties are uniform across the image, but this assumption is unlikely to be valid for large images. This study aims to characterize the spatial variation in atmospheric properties over a typical mid-latitude area (southern England), and to assess the errors that would result from applying a scene-based atmospheric correction to data collected under this variable atmosphere. Two key atmospheric properties – aerosol optical thickness (AOT) and precipitable water content (PWC) – are assessed over two clear days in June 2006, and results show an AOT range of approximately 0.1–0.5 and a PWC range of 1.5–3.0 cm. Radiative transfer modelling shows that errors in reflectance of up to 1.7 percentage points, and up to a 5% change in normalized difference vegetation index, can be caused by AOT variability, but that PWC variability has minimal effects. Sensitivity analyses also show that the high uncertainty of many data sources used to provide AOT values for atmospheric correction may also lead to significant errors in the resulting products. The spatial variability of the atmosphere cannot be ignored, and we are in need of operational, generic methods to perform a spatially variable atmospheric correction.

1. Introduction and background

Remotely sensed data are typically used to generate quantitative products which require a high degree of accuracy – for example, satellite sensor data are applied to estimate Global Climate Observing System Essential Climate Variables (GCOS 2004) such as snow cover (Hall, Riggs, and Salomonson 1995), sea-surface temperature (Brown et al. 1999), albedo (Wielicki et al. 2005), water vapour (Gao and Kaufman 2003), and net primary productivity (Running et al. 2004). To produce these variables, remotely sensed data must undergo atmospheric correction to remove the perturbing effects of the atmosphere (absorption and scattering of light by atmospheric constituents) from the data, and thus allow results to be determined accurately in physical units (Slater 1980). A range of atmospheric correction methods can be used with satellite sensor data but most studies assume that the atmosphere is spatially uniform across the image. Applying a spatially variable atmospheric correction is difficult in the absence of appropriate high-spatial resolution ground data and may be impossible when working with archived images, so most studies apply a scene-wide correction or use a generic atmospheric model such as MODTRAN (Moderate Resolution Atmospheric Transmission; Berk et al. 1999) or 6S (Vermote, Tanre, et al. 1997). However, over large images such as those from Landsat (185 km × 185 km) or the Disaster Monitoring Constellation (DMC; with a swath width
of 650 km), the atmosphere is likely to vary and so uniform correction methods may introduce significant errors in the resulting data products. This paper develops a previous study (Wilson, Milton, and Nield 2012) by further investigating errors associated with uniform atmospheric correction over large images of southern England. We first quantify the spatial variability of the atmosphere over southern England on a clear (cloud-free) day, and then assess the magnitude and range of errors associated with uniform atmospheric correction over this area, in terms of both radiance and normalized difference vegetation index (NDVI).

1.1. Background

In the early days of satellite remote sensing, simple scene-based atmospheric correction techniques such as Dark Object Subtraction (Chavez 1988) were generally used. In the 1990s and 2000s there was significant development in per-pixel approaches designed for use with hyperspectral imagery. However, there is a lack of true pixel-based correction methods for multispectral imagery. Tools such as ATCOR (Richter 2004) and FLAASH (Cooley et al. 2002) can be used to perform a pseudo-pixel-based correction of multispectral images, but the methods that are used for extracting spatially variable atmospheric information from multispectral imagery are limited. Thus, the majority of multispectral atmospheric corrections performed today are scene-based, using constant values of atmospheric parameters across the scene and not taking into account the spatial variability of the atmosphere over the image.

An example of an atmospheric correction system which takes into account spatial variability is the Landsat Ecosystem Disturbance Adaptive Processing System (LEDAPS; Masek et al. 2006), which estimates aerosol optical thickness (AOT) over areas of dense dark vegetation (DDV) in the image using the Kaufman et al. (1997) method. The AOT data are then interpolated to 1 km resolution and used to parameterize 6S (Vermote, El Saleous, et al. 1997) to perform the atmospheric correction. However, there are a number of problems with this method: (i) it does not take into account fine-scale variability in AOT; (ii) it can only estimate AOT over areas of DDV, thus making it impossible to use over areas without DDV, such as deserts; and (iii) it is only implemented for Landsat images because the AOT retrieval method requires the use of the Landsat shortwave infrared bands.

It is particularly important to quantify errors resulting from these scene-based atmospheric corrections due to a number of recent developments within remote sensing. First, the use of large images with a variety of spatial resolutions in environmental studies is becoming increasingly common, due both to the increased availability of large images and the policy-driven need for large area studies, particularly those relating to environmental change. Typical sensors include the Moderate Resolution Imaging Spectroradiometer (MODIS; 500 m resolution) and the DMC (30 m resolution), which both produce images that cover very large areas (a single DMC image can cover approximately half the area of England). Second, the wider availability of atmospheric correction as an option in standard image-processing software has made it more likely that such procedures will be used. Third, data obtained from quantitative analysis of remotely sensed images are now in widespread use for a variety of important scientific projects, and errors in these data could have serious consequences. In climate modelling, significant errors in input data caused by incorrect atmospheric correction could result in misleading predictions being reported to policymakers. For example, Saleska et al. (2007) stated that the Amazon rainforest was more resilient to short-term climatic fluctuations than previously thought (as shown by a significant increase in enhanced vegetation index), but Samanta et al.
(2010) showed that these inferences were due to the use of cloud- and aerosol-contaminated satellite data in the original study.

There has been discussion within the community as to whether atmospheric correction is required in all situations. Song et al. (2001) state that atmospheric correction is not required for applications which require only a single image and do not need the data to be in physical units. For example, they argue that performing a maximum likelihood classification of a single Landsat image using training samples derived from the image data per se would give exactly the same result with and without atmospheric correction. This is because a uniform atmospheric correction would simply apply the same correction to each pixel in each band, thus changing the mean of each land-cover class, but not altering the covariance of the classes. However, if a spatially variable atmospheric correction were to be performed the correction applied each pixel would be different (based upon the atmospheric conditions over that pixel), and thus the covariances of the classes would change. Similarly, if the image was acquired through a spatially variable atmosphere but a uniform atmospheric correction was performed there would be a different error for each pixel, which would affect the covariances of the classes. This does not just apply to classification: regardless of the final application of the image data, atmospheric correction is always required if the image was acquired under a spatially variable atmosphere, as each pixel will have undergone different atmospheric perturbation. In particular, band ratios such as NDVI can be significantly affected by the atmosphere due to the wavelength dependence of atmospheric scattering; ignoring the spatial variability of the atmosphere could cause apparent spatial trends in vegetation greenness, which would actually have been caused by a spatial trend in atmospheric conditions.

1.2. Atmospheric parameters of interest

The primary atmospheric constituents that affect remotely sensed measurements are mixed gases, ozone, aerosols, and water vapour. Concentrations of atmospheric mixed gases are controlled by atmospheric pressure, and ozone concentrations can be modelled effectively by latitude and season (Van Heuklon 1979). However, aerosol and water vapour concentrations vary significantly both spatially and temporally, and thus contemporaneous data on these must be provided when atmospherically correcting satellite images.

AOT, also known as aerosol optical depth (AOD), is a dimensionless measure of the degree to which aerosols restrict the transmission of light through the atmosphere, defined as the integrated extinction coefficient due to aerosols through a vertical column of unit area in the atmosphere (Iqbal 1983).

Water vapour in the atmosphere can be quantified in two ways: integrated water vapour (IWV), the vertically integrated mass of water per unit area (kg m$^{-2}$), or precipitable water content (PWC), the height of an equivalent column of liquid water (mm) (Iqbal 1983).

1.3. Previous work

Previous studies that have assessed AOT and PWC variability have typically used either (i) low-resolution data or (ii) daily, weekly, or monthly composites, which are relevant for climate-related studies but not for assessing the spatial variability in AOT and PWC at the specific instant of satellite image acquisition.

Steven and Rollin (1986) found a spatial variation in path radiance within some airborne images, suggested that this could have been caused by variation in the AOT,
and recommended caution when assuming atmospheric homogeneity for remote-sensing applications. More recently, González et al. (2003) and Koelemeijer, Homan, and Matthijsen (2006) examined the spatial variation of AOT across Europe using MODIS and ATSR-2 data, respectively. These data were averaged to monthly or yearly periods, and so only provide estimates of an average variability in AOT of 0.2–0.5. González et al. (2003) found a wide range in AOT values across Europe, with values of 0.5–0.6 in industrialized areas of Germany and northern Italy, and values of 0.1 in rural areas of France, Spain, and Norway, which suggests that local emissions are particularly important in determining AOT values. Koelemeijer, Homan, and Matthijsen (2006) also found significant local effects, with a number of cities easily distinguishable as peaks in the data, and particularly low AOT in mountainous areas. The AOT values in southern England from the same study reflect this, with high values around London and the Thames Estuary and generally low values in rural Cornwall. González et al. (2003) also found that AOTs can increase by up to 300% over relatively short distances (along a transect from Germany to Belgium), and similar gradients occurred in their data for the UK (e.g. an increase of 275% from east Kent to mid-Oxfordshire). AOTs also vary temporally, both diurnally (Smirnov et al. 2002) and over weekly periods (Bäumer and Vogel 2007). These anthropogenic variations, along with the prevailing meteorological situation, significantly impact the spatial variability of AOT, and thus yearly or monthly averages do not provide the information required for assessing the effect of uniform atmospheric correction procedures.

A number of studies have found significant variability in PWC, often over relatively short distances. Jedlovec (1990) measured PWC at 100 m resolution over the Tennessee Valley region of the USA and found significant variability in PWC at scales from a few kilometres to tens of kilometres. Furthermore, a significant change between the moist and dry parts of the study area (38 and 16 mm of PWC, respectively) occurred very suddenly over approximately 5 km. Over larger study areas, with lower-resolution data, Grody, Gruber, and Shen (1980) found a PWC range of approximately 25–65 mm across the Pacific Ocean (although the data were not cloud screened) and Holben and Eck (1990) found a coefficient of variation of less than 25% between stations located up to 500 km apart in the Sahel.

1.4. Study area and period

The study used data over southern England (the grey area in Figure 1) from 16 and 17 June 2006. These were typical mid-latitude clear days during the NCAVEO Field Campaign (Milton et al. 2011), when a range of ground and satellite data were available. The months of May and June tend to provide the best conditions for the acquisition of satellite data in the UK, due to reduced cloud cover and increased sky clarity. Furthermore, the mixed urban–rural mosaic land cover is representative of that found in many countries, and will allow the testing of satellite AOT retrieval algorithms under realistic conditions.

The meteorological situation changed significantly during these two days, as a high-pressure system migrated from the southern Atlantic Ocean over southern England to Germany. The passage of this weather system caused significantly different wind directions on the two days (with average directions, relative to north, of 287° on the 16th and 192° on the 17th). Field observations confirm that conditions on the 17th were more variable than on the 16th (Milton et al. 2011), with an increase in cloudiness and a reduction in sky clarity after 11:00 UTC.
2. Data sources and validations

AOT and PWC can be measured using a variety of ground instruments and satellite products with a range of spatial and temporal resolutions (Table 1). The uncertainty of these methods is determined by the location in which they are used, with factors such as land cover and aerosol type having a major effect, so it is important to perform validation for the study site.

Validation of satellite measurements against ground measurements is challenging for a number of reasons, including: (i) the lack of exactly coincident measurements, (ii) differences in cloud screening, (iii) different measurement variables, and (iv) the fundamental difference between areally integrated measures from satellites and point-based measurements from ground instruments. The Ichoku et al. (2002) spatio-temporal subset approach for validation is used here, comparing a spatial subset from the satellite data (5 × 5 pixels) to a temporal subset from ground measurements (±30 min). Ichoku et al. (2002) justified the size of these subsets based upon an estimate of average aerosol front speed, the requirement to obtain a statistically significant sample size, and the observation that larger window sizes could introduce errors from cloudy pixels and changing topography.

2.1. Aerosol robotic network sun photometry

Sun photometers estimate AOT and PWC based upon measurements of solar irradiance in a number of wavelengths. Here we use automatically cloud-screened data (Level 1.5) collected by the Aerosol Robotic Network (AERONET) Cimel CE-318 sun photometer situated at the Chilbolton Facility for Atmospheric and Radio Research (CFARR) (Holben...
Table 1. Summary of characteristics and accuracy of all data sources. The study area validation column provides the validation statistics in the form of root mean square errors (RMSEs) or Ichoku comparison $p$-values. The latter are considered significant, and thus shown in bold, when less than 0.05.

| Source                  | Type          | Spatial resolution | Temporal resolution          | Official validation | Study area validation | Closest acquisition time |
|-------------------------|---------------|--------------------|-------------------------------|---------------------|-----------------------|--------------------------|
| AERONET                 | Ground        | One location       | Every 15 min, in good weather | $\pm 0.02$          | –                     | 09:37                    |
| Met Office              | Ground        | 36 stations across study area | Hourly | – | RMSE: 0.05–0.47 (for visibility 40–10 km) | 10:00                    |
| MODIS AOT (MOD04)       | Satellite     | 10 km              | Daily merged or once per orbit | $\pm 0.05 \pm 0.15_r$ | $p < 0.3$             | 16th: 11:34              |
| GlobAerosol             | Multi-Satellite | 10 km              | Daily                        | RMSE: 0.12          | 16th: $p < 0.2$       | 17th: 10:38              |
| MODIS PWC (MOD05)       | Satellite     | 1 km               | Daily merged or once per orbit | RMSE: 1.7 mm        | 16th: $p < 0.5$       | 17th: 10:38              |
| BIGF Water Vapour       | Ground        | 25 stations across study area | Hourly | – | RMSE: 1.5 mm | 10:00                    |
et al. 1998). We used a simple time-for-space substitution to obtain an estimate of the spatial variability over the whole area from this single point measurement, by taking the AERONET measurements over the entire daylight period and assuming that they are representative of the AOT across the whole study area.

Sun photometers are used as reference data within this study, as they are currently the most accurate method for measuring AOT (Wang et al. 2009). Errors are low: approximately ±0.02 for AOT (Eck et al. 1999) and with a PWC RMSE of 2.9 mm (Liu et al. 2011).

The time-for-space assumption was examined by modelling the passage of aerosol particles across the UK during the study period using the Hybrid Single Particle Lagrangian Integrated Trajectory Model (HYSPLIT; Draxler and Rolph 2003), using gridded one degree-resolution meteorological data from the NCEP Global Data Assimilation System archive (Kalnay et al. 1996). The model was parameterized to simulate an ensemble of possible back-trajectories for a single particle located above the AERONET site at 19:00 UTC back to its starting location at 05:00 UTC (the start and end times of the AERONET data during the study period). Simulations were run for particles at heights of 500, 1000, and 1500 m to capture the differing trajectories produced by winds of variable height. These heights were chosen based upon the finding of Matthias et al. (2004) that 80–90% of the AOT is produced by aerosols in the planetary boundary layer, which was found to be at a height of 1204 ± 481 m at the Aberystwyth station, located approximately 100 km outside the study area.

A simple ‘contributing area’ for the AERONET site was then calculated as the concave hull of the resulting trajectories. These estimated areas for 16 and 17 June 2006 (Figure 1) show that the time-for-space substitution covered 23% and 19% of the study area on the 16th and 17th, respectively. The contributing area for each day was markedly different due to contrasting meteorological conditions, and a large proportion of the contributing area was outside of the study area (62% and 63%, respectively).

### 2.2. Met Office visiometry

AOT was estimated from hourly measurements of horizontal visibility (accurate to ±10%) acquired by a network of UK Met Office stations across the study area (UK Met Office 2006) using Koschmieder’s equation (Koschmieder 1925; Horvath 1981):

\[
V = \frac{3.912}{\tau},
\]

where \( V \) is the visibility in km and \( \tau \) is the AOT.

Koschmieder’s equation relates horizontal visibility and horizontal extinction coefficient measurements, but is now widely used for calculating vertical extinction coefficients (i.e. AOT) from horizontal visibility. This mixing of horizontal and vertical measurements relies on many assumptions which are often invalid (Chan 2009), and there are broader issues with the choice of coefficients in the equation (Middleton 1952; Horvath 1971, 1981). Previous studies comparing AOT and visibility-based estimates of AOT have found correlations ranging from 0.38 (So, Cheng, and Tsui 2005) to 0.89 (Chen et al. 2009). However, despite these limitations, visibility-based estimates of AOT are still useful due to their high spatial and temporal resolutions, as well as the wide availability of data collected according to World Meteorological Organization standards.
2.3. **MODIS AOT (MOD04)**

The MOD04 product from the MODIS sensors on the Terra and Aqua satellites provides AOT estimates at 10 km resolution using an algorithm based on shortwave infrared measurements and the use of a Radiative Transfer Model look-up table (Remer et al. 2006). The official MODIS validation report for the latest version of the algorithm (Collection 5.1) (Remer et al. 2006) states that 67% of the retrievals were within the expected uncertainty ($\pm 0.05 \pm 0.15 \tau$), which has been confirmed by independent validations (Levy et al. 2010). Results improve when only pixels with the highest quality assurance confidence are used, as in this study. Many assessments of MOD04 accuracy in the literature are based upon previous versions of the algorithm (Collection 4), but the current algorithm (Collection 5.1) has significantly improved the accuracy. The accuracy is seasonally variable (El-Metwally et al. 2010), probably due to the seasonal changes in aerosol types present over some of the sites used in ‘the validation study’ (or something similar).

Validation for the study area was performed between the MOD04 product and the AERONET site at Chilbolton using the method of Ichoku et al. (2002) (Figures 2(a) and (b)). Results from t-tests showed that there was no significant difference between the samples obtained from MODIS and AERONET ($p = 0.83$ and $p = 0.28$ for the 16th and 17th, respectively), and thus they are likely to have come from the same distribution of AOT values. An example image is shown in Figure 3(b).

2.4. **GlobAerosol**

The GlobAerosol product is produced by merging AOT products from the ATSR-2, AATSR, MERIS, and SEVERI sensors (Thomas et al. 2010) at 10 km resolution. Poor data are excluded based upon a number of checks, and merging is performed using temporal interpolation, with observations weighted by their error estimates (Siddans et al. 2007). The official validation against AERONET measurements found that the AATSR-derived data set was most accurate, with a RMSE of 0.07 (Poulsen et al. 2009). Although the merged product has a lower accuracy (Poulsen et al. 2009), the major advantage is that the merging process ensures higher spatial coverage of the area.

A comparison between the GlobAerosol merged product and the AERONET site at Chilbolton (Figures 2(c) and (d)) shows a smaller difference between AERONET and GlobAerosol measurements on 16 than on 17 June. Results from t-tests indicate that there was a significant difference between the AOT samples from MODIS and AERONET on the 17th ($p = 0.0008$), but no significant difference on the 16th ($p = 0.20$). An example image is shown in Figure 3(a).

2.5. **MODIS PWC (MOD05)**

The MOD05 product provides PWC estimates at 1 km resolution, based upon a ratio of adjacent bands with and without water absorption features. Official validation for the MODIS water vapour product is limited (Gao and Kaufman 2003), with RMSE based on a microwave radiometer data set of 1.7 mm corresponding to an approximate 5–15% error for the PWC range found over southern England (10–40 mm). Comparisons of the MOD05 PWC estimates to radiosonde and GPS-based measurements at Herstmonceux in southern England (50.889 N, 0.324 E) found a positive bias of 10% and 7%, respectively (Li, Muller, and Cross 2003), and comparisons in the Tibetan Plateau produced a similar result to the official validation (1.95 mm RMSE).
Figure 2. Boxplots showing the summary statistics for the validation of each satellite product against AERONET data.
Validation between the satellite data and the AERONET PWC measurements at Chilbolton (Figures 2(e) and (f)) indicated similar relationships between the spatial and temporal subsets on the 16th ($p = 0.437$), but not on the 17th ($p = 0.004$), as found with the other data sets. Again, the satellite data have a wider range, but on both days the AERONET data were encompassed within this range. An example image is shown in Figure 3(c).

2.6. GPS water vapour

Measurements of delays in GPS L-band radio signals passing through the atmosphere can be used to quantify the water vapour in the atmosphere above the GPS receiver (Bevis et al. 1992). The British Isles Continuous GNSS Facility (BIGF; NERC 2012) uses these
methods to provide estimates of IWV at all BIGF stations on an hourly basis. Previous validations of GPS-derived water vapour estimates against radiosonde and satellite data have generally produced errors of 1–2 mm (Becker et al. 2003; Wang et al. 2007; Wolfe and Gutman 2000; Tregoning et al. 1998). However, all the data used in these validations were processed from the GPS data by the authors, with specific parameterizations for the area of study. The BIGF product is a national operational product, and thus may be expected to have a lower accuracy.

A BIGF measurement site is co-located with the AERONET site at Chilbolton, and validation for all days with at least two matching measurements in the period August 2009 to November 2010 produced an average daily RMSE of 1.5 mm, with a maximum of 6.3 mm and a Pearson correlation coefficient of 0.976, showing a good agreement between GPS-derived and AERONET-derived estimates.

3. Simulation of uniform atmospheric corrections

The 6S radiative transfer model (Vermote, Tanre, et al. 1997) was used to simulate a uniform atmospheric correction over southern England, using the data on spatial variability described above. The Py6S (Wilson 2013) interface to 6S was used to allow hundreds of individual simulations to be run in an automated manner. Simulations were run in two stages: first to generate top-of-atmosphere (TOA) radiance from a representative vegetation spectrum under a given set of atmospheric parameters \( P_{up} \), and second to atmospherically correct the TOA radiance to ground reflectance under a different set of atmospheric parameters \( P_{down} \). \( P_{up} \) was set to the 5% or 95% quantile of the AOT or PWC values and \( P_{down} \) was set to the mean of the AOT or PWC values (from Tables 2 and 3), thus simulating the uniform atmospheric correction of a pixel measurement which was actually acquired in extreme conditions. Simulations were performed for Landsat bands 1–4, and 6S parameters other than AOT and PWC were set to appropriate values for southern England. Results from the simulations were retrieved as reflectance values. To assess the effect on a standard remote-sensing product, NDVI was also calculated from these reflectances.

The errors resulting from a uniform atmospheric correction are conceptually the same as those resulting from uncertainty in the atmospheric parameters: both are caused by differences between the true parameter value and that used for correction. Thus, a sensitivity analysis was also performed to assess the effects of the uncertainties in the data sources (as listed in Table 1) on remote-sensing products.

Table 2. Summary statistics showing the range of AOT values across southern England during the study period for each data source. Q05 and Q95 are the 5% and 95% quantiles, respectively.

| Source         | Min. | Max. | Mean | Q05 | Q95 |
|----------------|------|------|------|-----|-----|
| *(a)* 16 June 2006 |      |      |      |     |     |
| AERONET        | 0.120| 1.130| 0.291| 0.156| 0.464|
| Met Office     | 0.156| 0.391| 0.260| 0.156| 0.391|
| GlobAerosol    | 0.071| 0.496| 0.287| 0.155| 0.417|
| MODIS          | 0.078| 0.460| 0.258| 0.139| 0.398|
| *(b)* 17 June 2006 |      |      |      |     |     |
| AERONET        | 0.153| 0.436| 0.216| 0.160| 0.332|
| Met Office     | 0.145| 0.559| 0.296| 0.175| 0.489|
| GlobAerosol    | 0.050| 0.338| 0.164| 0.082| 0.258|
| MODIS          | 0.063| 0.440| 0.223| 0.090| 0.386|
4. Spatial variability over the study area

4.1. Aerosol optical thickness

The AOT range over the study area on 16 and 17 June 2006 was approximately 0.1–0.5 (Table 2). This is large, given that these measurements were acquired on days with mostly clear skies across the study area, and shows that there is higher spatial variability in AOT than that suggested by visual examination of sky conditions.

The overall range in AOT was similar for both days, but all data sources showed significantly higher variability on the 17th (Figure 4). This is consistent with the more changeable weather conditions on the 17th, as noted by the records from the NCAVEO Field Campaign (Milton et al. 2011). Similarly, the median values for each data set are very similar on the 16th, but not on the 17th. In all cases the satellite-based data sets (MODIS and GlobAerosol) had a lower minimum, which is likely to be due to errors in separating at-sensor radiance into ground reflectance and aerosol-scattering components.

The AOT data obtained from Met Office visibility measurements have a similar range to the other data sets on the 16th, but overestimate AOT on the 17th. This was probably caused by failure of the assumptions inherent in the visibility to AOT conversion due to the meteorological conditions. For example, local conditions can reduce horizontal

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Table 3. Summary statistics showing the range of PWC (in cm) across southern England during the study period for each data source. Q05 and Q95 are the 5% and 95% quantiles, respectively.

| Source     | Min. | Max. | Mean  | Q95  | Q05  |
|------------|------|------|-------|------|------|
| AERONET    | 1.433| 2.257| 1.859 | 2.236| 1.486|
| BIGF       | 1.750| 2.820| 2.138 | 2.565| 1.813|
| MODIS      | 0.212| 5.588| 2.180 | 2.772| 1.646|
| AERONET    | 1.892| 2.724| 2.433 | 2.688| 1.975|
| BIGF       | 1.820| 3.250| 2.676 | 3.190| 1.904|
| MODIS      | 1.458| 6.218| 2.763 | 3.379| 1.995|

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Figure 4. Boxplots showing the range of AOT values found over southern England during the study period according to each data source. (a) 16 June, (b) 17 June.
visibility at ground level, but not significantly affect the vertical extinction coefficient measured by AOT.

Generally the AERONET measurements have the lowest inter-quartile range (16th: 0.12; 17th: 0.04) due to the time-for-space substitution not capturing the variability across the entire study area, but have a number of high outliers (including a value of 1.13 on the 16th). These outliers are likely to be due to poor performance of the automated cloud-screening algorithm used for the level 1.5 data, which performs relatively poorly for large areas of temporally and spatially homogeneous cloud (Smirnov et al. 2000).

4.2. Precipitable water content

The range agreement between the data sources for PWC is weaker than for the AOT data sets (Figure 5), but an approximate range for PWC, taking into account expected values (Randel et al. 1996) and obvious outliers on the 16th, is approximately 1.5–3.0 cm on the 16th and around 2.0–3.5 cm on the 17th (Table 3). The larger values on the 17th were probably caused by the southerly/southwesterly winds bringing moist air masses from the Atlantic Ocean over the study area. Again, the satellite data had the largest range, with many outliers for MODIS on 16 June. High outliers may have been caused by incorrect cloud screening (as clouds will have a significantly higher PWC) but the very low values (a minimum of 0.2 cm for MODIS) may be plausible in certain areas (a maximum PWC of 3 mm was found in Norway by Mook 1978). The MODIS PWC data set has a significantly higher resolution than the AOT data sets (1 km compared with 10 km) and is likely to record more small-scale variation that would be averaged out in a lower-resolution data set, and thus to have a larger range. The BIGF data compare well with the other data sets, with a smaller range than MODIS but similar inter-quartile ranges, showing the utility of this relatively new measurement approach.

4.3. Summary

All data sources confirm that AOT and PWC over the study area were not uniform during the study period. As there were mainly clear skies during this time, it is likely that the

Figure 5. Boxplots showing the range of PWC values found over southern England during the study period according to each data source. (a) 16 June, (b) 17 June.
measured variability is a ‘best case’ scenario and that AOT variability will be greater in other situations. Thus, the assumption of atmospheric spatial uniformity made by atmospheric correction methods is not valid across a large area of southern England, and probably for other mid-latitude areas.

5. Effects of uniform atmospheric correction

We assume that a uniform correction was performed using the average AOT or PWC value over the study area. This would lead to correction of some pixels with a higher AOT than was actually present over the pixel, and others with a lower AOT than was actually present. Simulations were performed with the 5% and 95% quantiles of AOT and PWC, as this allows us to state that 10% of the pixels in the image will have an error of at least that found from the simulations.

5.1. Aerosol optical thickness

The results of the Radiative Transfer Model simulations show that atmospheric correction of data acquired under high AOT and corrected with lower AOT produces erroneously high reflectances (Table 4). This is due to the increased scattering caused by the aerosols that was not corrected by atmospheric correction. The error has a significant spectral dependence, with higher errors for lower wavelengths (blue) and very low errors for high wavelengths (NIR), and an overall range of 0.1–1.3 percentage points of reflectance. In this situation, NDVI decreased as the red reflectance increased relative to NIR reflectance. The absolute NDVI difference was low, but the error reached 5% for the Met Office data set on the 17 June. Even relatively small errors in NDVI may affect derived products such as estimates of biomass production; for example, Kaufman and Holben (1993) found that a NDVI difference of 0.04 corresponded to biomass production errors of 11–30%.

To put these errors in context, the Noise Equivalent Delta Radiance (NEΔL) and equivalent Noise Equivalent Delta Reflectance (NEΔρ) were calculated for each Landsat band under the simulation conditions, using the NEΔρ formula and data in Scaramuzza et al. (2004) (Table 5). The errors due to a uniform atmospheric (Table 4) are significant,

Table 4. Effects of a uniform atmospheric correction performed over an area with AOT variability from each data source. Values in the table are the reflectance differences for 95% perturbation; 5% perturbation, with reflectance values in percentages. Note that increases in AOT (using the 95% percentile of the AOT data from the data source) cause increased reflectance for all bands, but a decrease in NDVI.

| Source        | ρ_B  | ρ_G  | ρ_R  | ρ_NIR | NDVI       |
|---------------|------|------|------|-------|------------|
| (a) 16 June 2006 |      |      |      |       |            |
| AERONET       | +1.3 | −1.0 | +1.1 | −0.8  | +1.0; −0.8 | +0.1; −0.1 | −0.026; 0.027 |
| Met Office    | +1.0 | −0.8 | +0.8 | −0.6  | +0.7; −0.6 | +0.1; −0.1 | −0.017; 0.022 |
| GlobAerosol   | +1.0 | −1.0 | +0.8 | −0.8  | +0.7; −0.7 | +0.1; −0.1 | −0.018; 0.027 |
| MODIS         | +1.1 | −0.9 | +0.9 | −0.7  | +0.8; −0.7 | +0.1; −0.1 | −0.019; 0.024 |
| (b) 17 June 2006 |      |      |      |       |            |
| AERONET       | +0.9 | −0.4 | +0.7 | −0.4  | +0.6; −0.3 | +0.1; −0.1 | −0.014; +0.014 |
| Met Office    | +1.5 | −0.9 | +1.2 | −0.8  | +1.1; −0.7 | +0.1; −0.1 | −0.030; +0.025 |
| GlobAerosol   | +0.7 | −0.6 | +0.6 | −0.5  | +0.5; −0.5 | +0.1; −0.2 | −0.009; +0.017 |
| MODIS         | +1.2 | −1.0 | +1.0 | −0.8  | +0.9; −0.7 | +0.1; −0.2 | −0.022; +0.025 |
at almost 30-fold greater than the NEΔρ for bands 1 and 2, approximately 20-fold greater for band 3, and 3-fold greater for band 4.

5.2. Precipitable water content

The reflectance differences caused by PWC perturbations are significantly lower than those for AOT perturbations, with a maximum error of 0.5 percentage points in the NIR and 0.02 percentage points in the visible (Table 6). They have the opposite spectral dependence to the differences due to AOT, with low errors at short wavelengths but high errors at longer wavelengths. This is because most multispectral satellite bands are deliberately located away from areas of the spectrum that experience significant water absorption, but Landsat band 4 (NIR, 0.76–0.90 μm) covers a water absorption feature (Gao and Goetz 1990). Although there is a varying effect between the NIR and red bands, it is not as significant as with the AOT perturbations, and thus the NDVI differences are much lower (with a maximum of 0.007).

5.3. Sensitivity analysis

The sensitivity analysis (Figure 6) shows how errors in AOT propagate to the resulting NDVI values for a range of errors. Comparison of the results to the uncertainties of each data set (Table 1) shows that there is a serious problem in using these data sets to provide AOT values for use in atmospheric correction procedures. The NDVI changes resulting from the official error estimates for each of the AOT data sources (Table 7) generated errors ranging from 2% to 7% for all data sources except AERONET. The effect of the AERONET error on NDVI is acceptable at less than 1%. This suggests that only AERONET data should be used for parameterizing atmospheric correction models, but

Table 6. Effects of a uniform atmospheric correction performed over an area with PWC variability from each data source. Values in the table are the differences from the true results for 95% perturbation and 5% perturbation, and reflectance values are in percentages. ρ_B, ρ_G, ρ_R, and ρ_NIR are the reflectances in the Landsat blue, green, red, and NIR bands, respectively.

| Source   | ρ_B | ρ_G | ρ_R | ρ_NIR | NDVI |
|----------|-----|-----|-----|-------|------|
| (a) 16 June 2006 |
| AERONET  | 0.00; 0.00 | −0.02; +0.02 | −0.02; +0.02 | −0.37; +0.40 | −0.002; +0.002 |
| BIGF     | 0.00; 0.00 | −0.02; +0.01 | −0.02; +0.02 | −0.39; +0.32 | −0.002; +0.002 |
| MODIS    | 0.00; 0.00 | −0.03; +0.02 | −0.03; +0.02 | −0.49; +0.44 | −0.003; +0.006 |
| (b) 17 June 2006 |
| AERONET  | 0.00; 0.00 | −0.01; +0.02 | −0.01; +0.02 | −0.22; +0.43 | −0.001; +0.002 |
| BIGF     | 0.00; 0.00 | −0.02; +0.04 | −0.02; +0.04 | −0.41; +0.72 | −0.002; +0.004 |
| MODIS    | 0.00; 0.00 | −0.03; +0.03 | −0.03; +0.04 | −0.49; +0.70 | −0.003; +0.007 |
this research has also shown that a fully spatially variable correction is needed, and as AERONET sites are sparsely distributed this is not possible. The errors shown in this sensitivity analysis are a result of inaccuracies in the AOT data sources, and have no relationship to the uniform atmospheric correction issues on which our main study is focused. Therefore, these errors could occur in any atmospheric correction that uses visibility, MODIS, or GlobAerosol data to obtain the AOT input, regardless of whether the atmosphere is spatially uniform or variable.

5.4. Summary
Performing a uniform atmospheric correction for water vapour does not introduce unacceptable errors in reflectance or NDVI, with a maximum error of 0.7 percentage points and 0.6%, respectively. Further simulations have shown that even in areas with very high water vapour levels, such as tropical rainforests, NDVI values are unlikely to be

Table 7. Resulting error in NDVI caused by AOT uncertainties (according to the official validation) for each data source, for AOTs of 0.2 and 0.4.

| Data source   | AOT = 0.2 | AOT = 0.4 |
|---------------|-----------|-----------|
| AERONET       | 0.51      | 0.70      |
| MetOffice     | 3.37      | 7.06      |
| GlobAerosol   | 3.29      | 4.08      |
| MODIS         | 2.13      | 3.72      |
significantly affected by variability in water vapour unless that variability approaches 80% of the mean value.

In contrast, performing a uniform atmospheric correction over a spatially variable AOT distribution may cause errors in reflectance of up to 1.5 percentage points, and errors of 5% in NDVI values. Overall 5% of the pixels in the image may have a reflectance error of \( >+1.5 \) percentage points, and another 5% of the pixels may have an error of \( <-1.0 \) percentage points. To put this in context, 10% of the pixels in a Landsat image is approximately 3.8 million pixels, covering an area of approximately 3500 km².

6. Conclusions

The spatial variability of the atmosphere over southern England was investigated by acquiring data on AOT and PWC from a wide range of ground- and satellite-based sources on two clear days. All data sources except the AERONET network of ground-based sun photometers had high uncertainty, but it was possible to extract a range of AOT and PWC over the study area for each day of 0.1–0.5 and 1.5–3.0 cm, respectively. These ranges show that there is significant variation in these properties across this area.

The errors that would be caused by performing a uniform atmospheric correction over the study area were assessed through simulations using Py6S. These showed that ignoring the spatial variation in AOT when performing atmospheric corrections could cause errors in reflectance and NDVI of up 1.3 percentage points and 5%, respectively, but that ignoring spatial variation in PWC caused maximum errors of 0.7 percentage points and 0.6%, respectively (an acceptable error primarily due to the strategic location of multispectral sensor bands away from water absorption features).

The magnitude of error in reflectance and NDVI suggest that applying spatially uniform atmospheric corrections to images may introduce significant errors. Although some atmospheric correction tools take into account the variability of the atmosphere, these methods have significant limitations (for example, LEDAPS is only applicable to Landsat images). Thus, further research should take place to develop methods for performing spatially variable atmospheric corrections of a range of optical imagery. Developing software to apply a spatially variable atmospheric correction is relatively simple – the challenge is that no high-resolution data sets of AOT and PWC are currently available. Given that the results of this study show that spatial variability in AOT causes significantly higher errors than those from PWC, research efforts should focus on developing methods for retrieving high-resolution AOT from satellite images with an accuracy sufficiently high for these data to be used in atmospheric correction. Until these methods are available, currently available spatially variable atmospheric correction methods should be used where possible, and the spatial variability in atmospheric conditions should be investigated before using uniform atmospheric correction methods. Users should also be alert to the possibility of spatially variable errors in their results, caused by errors in the uniform atmospheric correction.

In conclusion, the results from this study show that there is significant variation in AOT and PWC across southern England during clear days. The variation in PWC is not significant in terms of the errors resulting from a uniform atmospheric correction, but ignoring the variation in AOT by performing a uniform atmospheric correction could cause significant errors (reflectance errors of over 20-fold NE\( \Delta \rho \), and NDVI changes \( >0.03 \)). The widespread availability of scene-based atmospheric correction procedures in
modern image-processing systems invites users to disregard spatial variability in the atmosphere and risks introducing significant errors into key derived products.

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References

Bäumer, D., and B. Vogel. 2007. “An Unexpected Pattern of Distinct Weekly Periodicities in Climatological Variables in Germany.” Geophysical Research Letters 34: L03819. doi:10.1029/2006GL028559.

Becker, M., M. Kirchner, P. Häfele, W. Söhne, and G. Weber. 2003. “Near Real-Time Tropospheric Signal Delay from EPN and German Permanent GPS Sites.” Proceedings of the EUREF Symposium, Toledo, June 4–7.

Berk, A., G. P. Anderson, L. S. Bernstein, P. K. Acharya, H. Dothe, M. W. Matthew, S. M. Adler-Golden, J. H. Chetwynd, S. C. Richtsmeier, B. Pukall, C. L. Allred, L. S. Jeong, and M. L. Hoke. 1999. “MODTRAN 4 Radiative Transfer Modeling for Atmospheric Correction.” Proceedings of SPIE 3756: 348–353.

Bevis, M., S. Businger, T. A. Herring, C. Rocken, R. A. Anthes, and R. H. Ware. 1992. “GPS Meteorology: Remote Sensing of Atmospheric Water Vapor Using the Global Positioning System.” Journal of Geophysical Research 97 (D14): 15787–15801. doi:10.1029/92JD01517.

Brown, O. B., P. J. Minnett, R. Evans, E. Kearns, K. Kilpatrick, A. Kumar, R. Sikorski, and A. Závody. 1999. MODIS Infrared Sea Surface Temperature Algorithm Theoretical Basis Document Version 2.0. Miami, FL: University of Miami 33149–1998.

Chen, L. X., B. Zhang, W. Q. Zhu, X. J. Zhou, Y. F. Luo, Z. J. Zhou, and J. H. He. 2009. “Variation of Atmospheric Aerosol Optical Depth and Its Relationship with Climate Change in China East of 100°E over the Last 50 Years.” Theoretical and Applied Climatology 96 (1–2): 191–199. doi:10.1007/s00704-008-0023-7.

El-Metwally, M., S. C. Alfaro, M. M. Abdel Wahab, A. S. Zakey, and B. Chatenet. 2010. “Seasonal and Inter-Annual Variability of the Aerosol Content in Cairo (Egypt) as Deduced from the Comparison of MODIS Aerosol Retrievals with Direct AERONET Measurements.” Atmospheric Research 97 (1–2): 14–25. doi:10.1016/j.atmosres.2010.03.003.

Gao, B., and A. F. H. Goetz. 1990. “Column Atmospheric Water Vapor and Vegetation Liquid Water Retrievals from Airborne Imaging Spectrometer Data.” Journal of Geophysical Research: Atmospheres (1984–2012) 95 (D4): 3549–3564. doi:10.1029/JD095iD04p03549.

Gao, B. C., and Y. J. Kaufman. 2003. “Water Vapor Retrievals Using Moderate Resolution Imaging Spectroradiometer (MODIS) Near-Infrared Channels.” Journal of Geophysical Research 108 (D13): 4389. doi:10.1029/2002JD003023.
GCOS. 2004. “Implementation Plan for the Global Observing Systems for Climate in Support of the UNFCCC.” Tech. Rep. GCOS-92. Accessed January 2013. http://www.wmo.int/pages/prog/gcos/Publications/gcos-138.pdf

González, C. R., M. Schaap, G. De Leeuw, P. J. H. Buitjies, and M. Van Loon. 2003. “Spatial Variation of Aerosol Properties over Europe Derived from Satellite Observations and Comparison with Model Calculations.” Atmospheric Chemistry and Physics 3: 521–533. doi:10.5194/acp-3-521-2003.

Grody, N. C., A. Gruber, and W. C. Shen. 1980. “Atmospheric Water Content over the Tropical Pacific Derived from the Nimbus-6 Scanning Microwave Spectrometer.” Journal of Applied Meteorology 19: 986–996. doi:10.1175/1520-0450(1980)019<0986:AWCOTT>2.0.CO;2.

Hall, D. K., G. A. Riggs, and V. V. Salomonson. 1995. “Development of Methods for Mapping Global Snow Cover Using Moderate Resolution Imaging Spectroradiometer Data.” Remote Sensing of Environment 54 (2): 127–140. doi:10.1016/0034-4257(95)00137-P.

Holben, B. N., and T. F. Eck. 1990. “Precipitable Water in the Sahel Measured Using Sun Photometry.” Agricultural and Forest Meteorology 52 (1–2): 95–107. doi:10.1016/0168-1923(90)90102-C.

Holben, B. N., T. F. Eck, I. Slutsker, D. Tanré, J. P. Buis, A. Setzer, E. Vermote, J. A. Reagan, Y. J. Kaufman, T. Nakajima, F. Lavenu, I. Jankowiak, and A. Smirnov. 1998. “AERONET—A Federated Instrument Network and Data Archive for Aerosol Characterization.” Remote Sensing of Environment 66 (1): 1–16. doi:10.1016/S0034-4257(98)00315-9.

Horvath, H. 1971. “On the Applicability of the Koschmieder Visibility Formula.” Atmospheric Environment (1967) 5 (3): 177–184. doi:10.1016/0004-6981(71)90081-3.

Horvath, H. 1981. “Atmospheric Visibility.” Atmospheric Environment (1967) 15 (10–11): 1785–1796. doi:10.1016/0004-6981(81)90214-6.

Ichoku, C., D. A. Chu, S. Mattoo, Y. J. Kaufman, L. A. Remer, D. Tanré, I. Slutsker, and B. N. Holben. 2002. “A Spatio-Temporal Approach for Global Validation and Analysis of MODIS Aerosol Products.” Geophysical Research Letters 29 (12): 8006. doi:10.1029/2001GL013206.

Iqbal, M. 1983. An Introduction to Solar Radiation. Toronto: Academic Press.

Jedlovec, G. J. 1990. “Precipitable Water Estimation from High-Resolution Split Window Radiance Measurements.” Journal of Applied Meteorology 29: 863–877. doi:10.1175/1520-0450(1990)029<0863:PWEFHR>2.0.CO;2.

Kalnay, E., M. Kanamitsu, R. Kistler, W. Collins, D. Deaven, L. Gandin, M. Iredell, S. Saha, G. White, J. Woollen, Y. Zhu, A. Leetmaa, R. Reynolds, M. Chelliah, W. Ebisuzaki, W. Higgins, J. Janowiak, K. C. Mo, C. Ropelewski, J. Wang, R. Jenne, and D. Joseph. 1996. “The NCEP-NCAR 40-Year Reanalysis Project.” Bulletin of the American Meteorological Society 77: 437–471. doi:10.1175/1520-0477(1996)077<0437:TNYRP>2.0.CO;2.

Kaufman, Y. J., and B. N. Holben. 1993. “Calibration of the AVHRR Visible and Near-IR Bands by Atmospheric Scattering, Ocean Glint and Desert Reflection.” International Journal of Remote Sensing 14 (1): 21–52. doi:10.1080/01431169308904320.

Kaufman, Y. J., A. E. Wald, L. A. Remer, B.-C. Gao, L. Li, and L. Flynn. 1997. “The MODIS 2.1-Mm Channel-Correlation with Visible Reflectance for Use in Remote Sensing of Aerosol.” IEEE Transactions on Geoscience and Remote Sensing 35 (5): 1286–1298. doi:10.1109/36.628795.

Koelemeijer, R. B. A., C. D. Homan, and J. Matthijsen. 2006. “Comparison of Spatial and Temporal Variations of Aerosol Optical Thickness and Particulate Matter over Europe.” Atmospheric Environment 40 (27): 5304–5315. doi:10.1016/j.atmosenv.2006.04.044.

Koschmieder, H. 1925. Theorie Der Horizontalen Sichtweite. Aachen: Kein & Nemnich.

Levy, R. C., L. A. Remer, R. G. Kleidman, S. Mattoo, C. Ichoku, R. Kahn, and T. F. Eck. 2010. “Global Evaluation of the Collection 5 MODIS Dark-Target Aerosol Products over Land.” Atmospheric Chemistry and Physics 10 (21): 10399–10420. doi:10.5194/acp-10-10399-2010.

Li, Z., J. P. Muller, and P. Cross. 2003. “Comparison of Precipitable Water Vapor Derived from Radiosonde, GPS, and Moderate-Resolution Imaging Spectroradiometer Measurements.” Journal of Geophysical Research 108: 4651. doi:10.1029/2003JD003372

Liu, Z., M. S. Wong, J. Nichol, and P. W. Chan. 2011. “A Multi-Sensor Study of Water Vapour from Radiosonde, MODIS and AERONET: A Case Study of Hong Kong.” International Journal of Climatology 33: 109–120.

Masek, J. G., E. F. Vermote, N. E. Saleous, R. Wolfe, F. G. Hall, K. F. Huemmrich, F. Gao, J. Kutler, and T. Lim. 2006. “A Landsat Surface Reflectance Dataset for North America, 1990–2000.” IEEE Geoscience and Remote Sensing Letters 3 (1): 68–72. doi:10.1109/LGRS.2005.857030.
Steven, M. D., and E. M. Rollin. 1986. “Estimation of Atmospheric Corrections from Multiple Aircraft Imagery†.” International Journal of Remote Sensing 7 (4): 481–497. doi:10.1080/01431168608954704.

Thomas, G., C. Poulsen, R. Siddans, E. Carboni, A. Sayer, and D. Grainger. 2010. “The Globaerosol Dataset: Using a Multi-Instrument Satellite Aerosol Dataset.” In EGU General Assembly Conference Abstracts, Vol. 12(1081), Vienna, May 2–7. Vienna: Copernicus.

Tregoning, P., R. Boers, D. O’Brien, and M. Hendy. 1998. “Accuracy of Absolute Precipitable Water Vapor Estimates from GPS Observations.” Journal of Geophysical Research 103: 28701–28710. doi:10.1029/98JD02516.

UK Met Office. 2006. “MIDAS Land Surface Stations Data (1853–Current) at NCAS British Atmospheric Data Centre.” Accessed August 2012. http://badc.nerc.ac.uk/view/badc.nerc.ac.uk_ATOM_dataent_ukmo-midas

Van Heuklon, T. K. 1979. “Estimating Atmospheric Ozone for Solar Radiation Models.” Solar-Energy 22 (1): 63–68. doi:10.1016/0038-092X(79)90060-4.

Vermote, E. F., N. El Saleous, C. O. Justice, Y. J. Kaufman, J. L. Privette, L. Remer, J. C. Roger, and D. Tanre. 1997. “Atmospheric Correction of Visible to Middle-Infrared EOS-MODIS Data over Land Surfaces: Background, Operational Algorithm and Validation.” Journal of Geophysical Research: Atmospheres (1984–2012) 102 (D14): 17131–17141. doi:10.1029/97JD00201.

Vermote, E. F., D. Tanre, J. L. Deuze, M. Herman, and J. J. Morcette. 1997. “Second Simulation of the Satellite Signal in the Solar Spectrum, 6S: An Overview.” IEEE Transactions on Geoscience and Remote Sensing 35 (3): 675–686. doi:10.1109/36.581987.

Wang, J., L. Zhang, A. Dai, T. Van Hove, and J. Van Baelen. 2007. “A Near-Global, 2-Hourly Data Set of Atmospheric Precipitable Water from Ground-Based GPS Measurements.” Journal of Geophysical Research 112 (D11): D11–107. doi:10.1029/2006JD007529.

Wang, Y., A. I. Lyapustin, J. L. Privette, J. T. Morisette, and B. Holben. 2009. “Atmospheric Correction at AERONET Locations: A New Science and Validation Data Set.” IEEE Transactions on Geoscience and Remote Sensing 47 (8): 2450–2466. doi:10.1109/TGRS.2009.2016334.

Wielicki, B. A., T. Wong, N. Loeb, P. Minnis, K. Priestley, and R. Kandel. 2005. “Changes in Earth’s Albedo Measured by Satellite.” Science 308 (5723): 825–825. doi:10.1126/science.1106484.

Wilson, R. T. 2013. “Py6s: A Python Interface to the 6S Radiative Transfer Model.” Computers and Geosciences 51: 166–171. doi:10.1016/j.cageo.2012.08.002.

Wilson, R. T., E. J. Milton, and J. Nield. 2012. “Spatial Variability of the Atmosphere across Southern England and the Resulting Error in Assuming a Uniform Atmospheric Correction.” Remote Sensing and Photogrammetry Society Annual Conference, Greenwich, September.

Wolfe, D. E., and S. I. Gutman. 2000. “Developing an Operational, Surface-Based, GPS, Water Vapor Observing System for NOAA: Network Design and Results.” Journal of Atmospheric and Oceanic Technology 17 (4): 426–440. doi:10.1175/1520-0426(2000)017<0426:DAOSBG>2.0.CO;2.