Algorithm for the retrieval of columnar water vapor from hyperspectral remotely sensed data

Alessandro Barducci, Donatella Guzzi, Paolo Marcoionni, and Ivan Pippi

A new algorithm for the retrieval of columnar water vapor content is presented. The proposed procedure computes the area of the $\text{H}_2\text{O}$ absorption centered about 940 nm to allow its integrated columnar abundance as well as its density at ground level to be assessed. The procedure utilizes the HITRAN 2000 database as the source of $\text{H}_2\text{O}$ cross-section spectra. Experimental results were derived from radiometrically calibrated hyperspectral images collected by the Airborne Visible–Infrared Imaging Spectrometer (AVIRIS) sensor over the Cuprite mining district in Nevada. Numerical simulations based on the MODTRAN 4 radiative transfer code were also employed for investigating the algorithm’s performance. An additional empirical $\text{H}_2\text{O}$ retrieval procedure was tested by use of data gathered by the VIRS-200 imaging spectrometer. © 2004 Optical Society of America

OCIS codes: 010.1280, 010.1320, 280.0280, 300.1030, 300.3700.

1. Introduction

Precise knowledge of atmospheric transmission and scattering is gaining growing relevance for obtaining accurate surface reflectance spectra from aerospace measurements. Retrieval of water vapor abundance is often used to remove the related absorption features from the radiance reaching the sensor and even for investigating the greenhouse effect and the energy balance on the ground. Aerospace remote sensing of atmospheric water vapor content may be performed with a passive microwave sensor for which the retrieval algorithms require data gathered over water surfaces; with a thermal infrared sensor operating over land under clear-sky conditions, a circumstance in which retrieval accuracy is limited by the assumed surface temperature and emissivity; or by measurement of the solar radiation reflected by a land surface in the visible and near-infrared spectral regions. In fact the retrieval of water vapor content over inland waters and the open sea shows strong inaccuracies because of the low reflectance of these waters, which causes the measured signal to be due mainly to aerosol scattering. Measurements over clouds, which show small spectral variation in reflectance, are hindered by changes in the penetration depth of the incoming solar radiation.

Following a spectroscopic approach, we developed a new mathematical method for quantitative retrieval of vertically integrated water vapor content, largely in the lower part of the troposphere. The algorithm was used to process both radiance spectra simulated with the MODTRAN4 radiative transfer code and hyperspectral images collected with the Airborne Visible-Infrared Imaging Spectrometer (AVIRIS) over the Cuprite mining district in southwest Nevada. In Section 2 we discuss the details of the proposed mathematical method and an additional empirical method to perform $\text{H}_2\text{O}$ retrievals from laboratory measurements obtained with the VIRS-200 imaging spectrometer. In Section 3 experimental results are compared with numerical simulations. Section 4 is devoted to conclusions and plans for future developments.

2. General Description of the Algorithm

The radiance measured by a downlooking imaging spectrometer can be written as the sum of the atmospheric radiance (path radiance) and the directly transmitted radiance that emerge from the observed target. In the visible and near-infrared parts of the electromagnetic spectrum the measured radiance
Fig. 1. Vertical atmospheric transmittance versus wavelength for three water vapor amounts (MODTRAN 4 simulations). The curves correspond to relative differences in the water vapor abundance amounts of 0.5, 1.0, and 2.0 with respect to the standard MODTRAN 4 value (corresponding to 1.46, 2.92, and 5.84 g cm\(^{-2}\), respectively). The simulation refers to a midlatitude summer atmosphere with a visibility of 20 km and a rural aerosol model. Note the presence of the O\(_2\) absorption band at 760 nm.

\[ L(k) = \int L_s(\lambda, \theta) S_\lambda(\lambda) \, d\lambda \]

\[ = \int \left[ L_{\text{path}}(\lambda, \theta) + \rho(\lambda, \theta) \frac{E(\lambda, \zeta) \cos \theta}{\pi} \right] \times \exp\left[-\tau(\lambda) \sec \theta\right] S_\lambda(\lambda) \, d\lambda, \]

\[ k = 1, \ldots, N, \]

where \( L_s(\lambda, \theta) \) is the at-sensor spectral radiance, \( L_{\text{path}}(\lambda, \theta) \) is the atmospheric path radiance, \( E(\lambda, \zeta) \) is the at-ground irradiance, \( \tau(\lambda) \) is the optical thickness of the plane-parallel atmospheric slab from ground to sensor, \( \rho(\lambda, \theta) \) is the surface reflectance, \( \theta \) is the viewing direction (the angle between the nadir and the viewing direction), \( \zeta \) is the zenith distance of the Sun, and \( S_\lambda(\lambda) \) is the spectral sensitivity of the \( k \)th spectral channel. It is clear that the radiance that reaches the sensor is modulated by the atmospheric transparency \( \exp\left[-\tau(\lambda)\right] \); thus it contains information pertaining to scattering particles (molecules and aerosols) and absorbing gas.

As suggested from previous research\(^{7,8} \) the atmospheric transparency depends strongly on the number of water vapor molecules found in the viewing path, a circumstance that allows us to retrieve the integrated water vapor abundance from the shape of one of its absorption bands.

By means of the MODTRAN4 radiative transfer code\(^{9,10} \) we simulated a standard atmospheric transparency for three water vapor amounts, as shown in Fig. 1. The strong dipole moment and the light hydrogen atoms that compose a water vapor molecule result in strong and broad absorption bands. The spectra that we obtained indicate that, for typical atmospheric conditions, the in-band transparency near 940 and 1140 nm is sensitive to changes in the integrated amount of water vapor, whereas the bands at 1380 and 1880 nm are often saturated. Another useful absorption band is that which is centered at 820 nm, but the correspondingly small change in transparency with water vapor content should be measured with higher accuracy (see Fig. 2).

To retrieve columnar amounts of water vapor from remotely sensed radiance acquired by hyperspectral sensors it is necessary to adopt some inverse modeling of atmospheric optical parameters. Most available models\(^{11,12} \) represent judicious compromises between a faithful representation of a physical system and its mathematical tractability. For example, the pressure broadening of a diatomic molecule at moderately low temperatures may be locally represented by the Elsasser (or regular) band model\(^{13-15} \); an infinite array of equally spaced spectral lines of equal strength and identical shape (identical half-widths). The distribution of energy levels for an asymmetric molecule, however, requires a statistical (or random) model, in which the lines are assumed to be randomly located. One can obtain a more realistic representation by assuming that the line intensities are governed by a probability-density function. Then the properties of the band will be determined by a weighted average of the single-line properties.

The main atmospheric gases are N\(_2\) (78%), O\(_2\) (21%), H\(_2\)O, Ar, CO\(_2\), and O\(_3\). In the visible part of the electromagnetic spectrum, transparency is affected mainly by ozone and aerosol absorption below 400 nm and by Rayleigh and Mie scattering. Here we want to focus our attention only on the water vapor absorp-
tion band located at 940 nm. The total water vapor transmission may be expressed as\(^1\),\(^2\) 

\[
\Gamma_\ell(\lambda, T) = \Gamma_0(\lambda, T) \Gamma_e(\lambda, T) \Gamma_i(\lambda, T) 
= \Gamma_e(\lambda, T) \Gamma_i(\lambda, T) 
. \tag{2}
\]

Here \(\Gamma_i(\lambda, T)\) is the percentage of radiation attenuated by water vapor single lines.\(^1\),\(^2\) \(\Gamma_e(\lambda, T)\) is the transparency that is due to water vapor continuum absorption, and \(T\) is the temperature.

Various mechanisms to explain the water vapor continuum absorption, including contributions from the extreme wings of the strong rotational water vapor lines in the far infrared as well as possible \((H_2O)_2\) dimer contributions, have been proposed. Generally, two contributions are recognized: one is due to a self-broadening term and the other, the so-called foreign broadened component, is due to the collisions by other atmospheric gases (mainly \(N_2\), which comprises 78\% of the total).

The first factor on the right-hand side of Eq. (2) takes into account the strong dependence of the water vapor absorption coefficient on temperature and on partial pressure in the atmosphere. The corresponding absorption coefficient \(k_c(\lambda, T, e, p_0)\) is modeled for a mixture of \(H_2O\) in \(N_2\) as follows\(^2\)–\(^3\):

\[
k_c(\lambda, T, e, p_0) = C^e_v(\lambda, T) w [e + (p_0 – e) \beta(T)] 
. \tag{3}
\]

where \(C^e_v(\lambda, T)\) is the self-broadening coefficient [molecules\(^{-1}\) cm\(^2\) atm\(^{-1}\)] for water vapor, \(w\) is the density [molecules cm\(^{-3}\)] of water vapor molecules, \(e\) is the water vapor partial pressure [atm], \(p_0\) [atm] is the total sample pressure, and \(\beta(T)\) is the ratio of the \(N_2\)-broadening coefficient to the self-broadening. This implies that for \(e \geq 0.1\) atm and \(p_0 = 1\) atm more than 90\% of the contribution to \(k_c(\lambda, T, e, p_0)\) comes from the self-broadening term and that \(k_c(\lambda, T, e, p_0)\) varies approximately as \(e^2\).

In the visible–near-infrared spectral range, considering water vapor concentrations of 0.4 to 4.0 g cm\(^{-2}\), the continuous absorption coefficient introduced above is far below the single-line absorption coefficient. In fact, single line absorption is the most important contribution to atmospheric absorption in the 940-nm band, and it is rather insensitive to temperature and pressure profiles in the atmosphere.

The monochromatic atmospheric transparency for a vertical path between the ground \((z = 0)\) and a final height \(z\) can be written as\(^2\)–\(^4\)

\[
\Gamma_i(\lambda) = \exp(-\tau(\lambda, z)) 
. \tag{4}
\]

where optical thickness \(\tau(\lambda, z)\) is expressed as

\[
\tau(\lambda, z) = \int_0^z \beta_{ext}(\lambda, \xi) d\xi = \int_0^z s(\lambda) n(\xi) d\xi 
, \tag{5}
\]

where \(\beta_{ext}(\lambda, \xi)\) represents the spectral extinction coefficient, \(n(\xi)\) is the density of the absorber, and \(s(\lambda)\) is the absorber’s cross section.

The basic idea of the proposed algorithm is to use the normalized line depth [see Eq. (7) below] to estimate the line area by spectral integration. The line area is a relevant parameter from any absorption band because it measures the total energy removed from the radiation field as a result of the spectral property considered. This parameter can be reliably estimated also from the radiance spectra acquired at moderate spectral resolution; moreover, the required in-band spectral integration helps to mitigate the effects of any uncorrelated noise source.

Let us consider a remote source that emits a spectral radiance \(L_s(\lambda, \theta)\) that is not affected by continuum as well as line absorption. At-sensor radiance
The basic idea of the proposed algorithm is to use residual line intensity $\eta(\lambda)$ that corresponds to the water vapor absorption band centered at 940 nm. This value is used for estimating the line area by spectral integration. The line-area estimation requires knowledge of the ideal spectral radiance as dimmed by continuous absorption only, $L_c(\lambda)$, at any $\lambda$ within the line. This value is obtained by use of samples of the observed spectral radiance $L_s(\lambda, \vartheta)$ near the wings of the line at a wavelength where line absorption becomes negligible. At least two samples, one on the blue flank and the other on the red flank, are selected to permit the fitting of an interpolation straight line. The interpolated radiance is then used as an estimate of continuum intensity. The straight line holds two nearby negligible. At least two samples, one on the blue flank and the other on the red flank, are selected to permit the fitting of an interpolation straight line. The interpolated radiance is then used as an estimate of continuum intensity. The straight line holds two nearby nonabsorption channels, and the curves with triangle symbols indicate a radiance spectrum simulated by a MODTRAN4 radiative transfer code for a standard midlatitude summer atmosphere for an observer at 1.5 km over the ground.

$L_o(\lambda, \vartheta)$ is related to emitted radiance $L_c(\lambda, \vartheta)$ as follows:

$$L_o(\lambda, \vartheta) = L_c(\lambda, \vartheta) \Gamma_o(\lambda) \Gamma_f(\lambda, \vartheta).$$  \hfill (6)

The normalized line depth $\eta(\lambda)$ at wavelength $\lambda$ can be reliably estimated as

$$\eta(\lambda) = \frac{L_o(\lambda, \vartheta) \Gamma_o(\lambda) - L_o(\lambda, \vartheta)}{L_c(\lambda, \vartheta) \Gamma_f(\lambda, \vartheta)} = 1 - \Gamma_f(\lambda, \vartheta),$$  \hfill (7)

which yields the fraction of monochromatic power absorbed by the atmospheric slab.

In Fig. 3 we schematically depict the measurement of the line area. $L_o(\lambda, \vartheta) \Gamma_o(\lambda)$ represents the expected continuum intensity; thus normalized line depth $\eta(\lambda)$ no longer depends on emitted radiance $L_o(\lambda, \vartheta)$ or on continuous absorption $\Gamma_f(\lambda, \vartheta)$.

Assuming that the shape of a single spectral line for a homogeneous path in a single absorbing gas is represented by the Lorentz profile, line depth $\eta(\lambda)$ integrated on spectral window $\Delta \lambda$ is given by

$$A = \int_{\Delta \lambda} \eta(\lambda) d\lambda = \int_{\Delta \lambda} \left[ 1 - \exp \left( -\frac{1}{\pi} \int_S T(\xi) \right) \right] S[T(\xi)] \times \frac{\delta \kappa[e(\xi)]}{(\kappa - \kappa_0)^2 + \{\delta \kappa[e(\xi)]\}^2} n(\xi) d\xi,$$  \hfill (8)

Here $\gamma$ is the ray path and

$$\frac{S[T(\xi)]}{\pi} \frac{\delta \kappa[e(\xi)]}{(\kappa - \kappa_0)^2 + \{\delta \kappa[e(\xi)]\}^2}$$

is the parametric expression of the line profile (Lorentz shape), which depends on line strength $S$, line center frequency $\kappa_0$, and the half-width at half-maximum $\delta \kappa$ that is related to collision broadening and Einstein’s spontaneous emission. The line strength depends mainly on gas temperature $T$, which in turn changes with height in the atmosphere.\(^{35}\) Line-width parameter $\delta \kappa$ instead is a function of altitude through the water vapor’s partial pressure $e$ being thermodynamic Doppler broadening negligible. To account for the whole spectral band we have to sum all single-line contributions blended in the observed spectral signature:

$$A \approx \int_{\Delta \lambda} \left[ 1 - \exp \left( -\frac{S}{\pi} \sum_{j=1}^{M} \frac{\delta \kappa[e(\xi)]}{(\kappa_j - \kappa_0)^2 + \{\delta \kappa[e(\xi)]\}^2} \right) \right] d\kappa,$$  \hfill (9)

where we have considered in the sum the total contribution from $M$ single lines, each one described by the Lorentz profile of Eq. (8). The unknown $\bar{N} = \int n(\xi) d\xi$ is the integrated amount of absorbers, namely, the mass of absorbers in an atmospheric column of uni-
tary cross section. It is worth noting that in expression (9) we have obtained an approximate relationship that may even depend on the assumption of single-line profiles that are independent of height. The assumption of line profiles that are independent of height holds true only for uniform atmosphere in which temperature and pressure do not change with altitude. However, this circumstance is partially verified in our case because we are considering airborne observations of the Earth’s surface from a relative height (above the ground) that was a limited range (3–20 km). The approximate result of expression (9) can also be accepted for unsaturated lines, in which case the exponent on the right-hand side of Eq. (8) can be expanded as a power series stopped to the first order. One may compute the double integral (in frequency k and position \( \xi \)) thus obtained by carrying out the frequency integral as a first step. As a result of this first integration the dependence on \( \delta k \) of the line profile is canceled out and an outcome proportional to \( S n(\xi) \) is obtained. We point out that line strengths \( S \) of transitions embedded in the near-infrared water-vapor band under consideration are almost insensitive to the small temperature variations encountered in the atmosphere when the height is changed within the aforementioned range. In fact, \( S \) is regulated mainly by population distribution in the two levels of the molecular transition. Population ratio \( P \) is governed by Boltzmann’s law:

\[
P = \frac{g_1}{g_0} \exp\left(-\frac{\Delta E_{10}}{kT(\xi)}\right),
\]

where, as can be shown, transition energy \( \Delta E_{10} \) is far above typical atmospheric \( KT(\xi) \) values at any height \( \xi \). Hence the exponential term is estimated in its asymptotic region where its \( T \) derivative is vanishing, thus making line strength \( S \) constant with height.

For typical airborne remote-sensing applications all the above reasons and approximations are verified at least partially, thus giving likelihood to the approximate relationship in expression (9).

If the chosen spectral interval \( \Delta \lambda \) covers the extension of the absorption band, then one can equivalently take the limits of integration of expression (9) from \(-\infty \) to \(+\infty \) without introducing significant error. A typical cross-section spectrum for the \( \text{H}_2\text{O} \) molecule (derived from the HITRAN 2000 database\(^{10} \)) is shown in Fig. 4.

Expression (9) represents the fundamental relationship for our retrieval algorithm: using Eq. (7), we compute a line-area estimate from experimental data (hyperspectral image). This estimate is then compared with the theoretical expectation of expression (9); thus an optimal estimate of the integrated amount of water vapor, \( N \), can be inferred. Figure 5 shows the main structure of our model. Using Eq. (7) for line-area estimation requires knowledge of the ideal spectral radiance \( L_{\nu}(\lambda, \delta) \Gamma_{\nu}(\lambda) \) as dimmed by the continuous absorption only, at any \( \lambda \) within the line. This end is achieved by use of the samples of the observed spectral radiance near the shoulders of the involved absorption line at wavelengths where the line absorption becomes negligible. At least two samples, one on the blue flank and the other on the red flank, are selected, thus allowing us to fit a suitable interpolation line. The interpolated radiance is then used as an estimate of continuum intensity. Let us note that this approach is markedly different from the standard channel-rationing techniques that are employed, for example, in the Moderate-Resolution Imaging Spectroradiometer (MODIS) algorithm theoretical base document.\(^{36} \) The main advantage of our algorithm is that it does not require any lookup tables to generate columnar amounts of water vapor because it numerically finds the best
3. Simple Modeling

The main drawback of our model for H$_2$O integrated abundance computation is the requirement of high-resolution cross-section spectrum as well as the implicit calculation burden. To arrange for a faster although approximate estimate, we derived from the original algorithm a simpler procedure, hereinafter called the simple model.

In Section 2 we demonstrated that the line area is a relevant quantity for the retrieval of water vapor abundance. One step of our method was the integration of optical thickness $\tau_{H_2O}(\lambda, z)$, or a related quantity, over a certain spectral interval $\Delta\lambda_{H_2O}$ where the H$_2$O absorption feature is localized. If we assume that $\tau_{H_2O}(\lambda, z)$ is a separable function of spectral $\lambda$ and spatial $z$ variables, it is easily to find that

$$\int_{\Delta\lambda_{H_2O}} \tau_{H_2O}(\lambda, z) d\lambda = \int_{\Delta\lambda_{H_2O}} s_{H_2O}(\lambda) d\lambda \int_{\gamma_1} n_{H_2O}(\xi) d\xi,$$

where the symbols are defined as in Sections 1 and 2.

We can also assume that the second integration on the right-hand side of expression (10) is expressed as the product of two amounts: the first is related to the columnar abundance $\bar{N}_{H_2O}$ of water vapor and the second one concerns the way in which the water vapor concentration is distributed in path length $\gamma_1$:

$$\int_{\gamma_1} n_{H_2O}(\xi) d\xi = \bar{N}_{H_2O} \int_{\gamma_1} f_{H_2O}(\xi) d\xi.$$

This calculation can be repeated for a different absorption line of any other atmospheric constituent, for instance O$_2$, that is a gas at constant mixing ratio, as are CO$_2$, N$_2$O, CO, and CH$_4$:

$$\int_{\Delta\lambda_{O_2}} \tau_{O_2}(\lambda, z) d\lambda = \int_{\Delta\lambda_{O_2}} \beta_{O_2}(\lambda) d\lambda \int_{\gamma_1} n_{O_2}(\xi) d\xi = \bar{N}_{O_2} \int_{\Delta\lambda_{O_2}} \beta_{O_2}(\lambda) d\lambda \int_{\gamma_1} f_{O_2}(\xi) d\xi.$$

Now we define $\varphi_{\text{mis}}$ as the line-area ratio for the two absorbers considered here, which can be computed starting from the measurements

$$\varphi_{\text{mis}} = \frac{\int_{\Delta\lambda_{H_2O}} \tau_{H_2O}(\lambda, z) d\lambda}{\int_{\Delta\lambda_{O_2}} \tau_{O_2}(\lambda, z) d\lambda} = \frac{(\bar{N}_{H_2O})_{\text{mis}} \int_{\Delta\lambda_{H_2O}} s_{H_2O}(\lambda) d\lambda \int_{\gamma_1} f_{H_2O}(\xi) d\xi}{\bar{N}_{O_2} \int_{\Delta\lambda_{O_2}} s_{O_2}(\lambda) d\lambda \int_{\gamma_1} f_{O_2}(\xi) d\xi}.$$

Fig. 5. Flow diagram showing the main steps of the retrieval algorithm. Starting from radiometrically corrected data (spectral radiance), the 940-nm line area is computed according to Eq. (7) (see text). Then the line area is estimated from the theoretical expectation of expression (9); an H$_2$O cross-section spectrum is introduced from the HITRAN 2000 database. Finally the two line areas are compared by numerical analysis and the vertically integrated water vapor abundance is computed.
The same ratio \( \Phi_{\text{mod}} \) can be predicted from a suitable theoretical model of atmospheric radiative transfer, yielding in ray path \( \gamma_2 \)

\[
\Phi_{\text{mod}} = \frac{\int_{\Delta H_2} \tau_{H_2}(\lambda, z) d\lambda}{\int_{\Delta O_2} \tau_{O_2}(\lambda, z) d\lambda}
\]

\[
(\tilde{N}_{H_2}O)_{\text{mod}} \int_{\Delta H_2} \beta_{H_2}(\lambda) d\lambda \int_{\gamma_2} f_{H_2}(\xi) d\xi
\]

\[
= \frac{(\tilde{N}_{H_2}O)_{\text{mod}} \int_{\Delta H_2} \beta_{O_2}(\lambda) d\lambda \int_{\gamma_2} f_{O_2}(\xi) d\xi}{\tilde{N}_{O_2} \int_{\Delta O_2} \beta_{O_2}(\lambda) d\lambda \int_{\gamma_2} f_{O_2}(\xi) d\xi}
\]

(14)

Let us note that, because \( O_2 \) has a fixed mixing ratio, the amount \( \tilde{N}_{O_2} \) remains constant even if the path length is changed. The ratio between Eqs. (13) and (14) reduces to

\[
\frac{\Phi_{\text{mis}}}{\Phi_{\text{mod}}} = \frac{\int_{\Delta H_2} \tau_{H_2}(\lambda, z) d\lambda}{\int_{\Delta O_2} \tau_{O_2}(\lambda, z) d\lambda}
\]

\[
(\tilde{N}_{H_2}O)_{\text{mis}} \int_{\gamma_1} f_{H_2}(\xi) d\xi \int_{\gamma_2} f_{O_2}(\xi) d\xi
\]

\[
= \frac{(\tilde{N}_{H_2}O)_{\text{mod}} \int_{\gamma_2} f_{H_2}(\xi) d\xi \int_{\gamma_1} f_{O_2}(\xi) d\xi}{(\tilde{N}_{H_2}O)_{\text{mod}} \int_{\gamma_2} f_{O_2}(\xi) d\xi}
\]

(15)

If the two ray paths coincide,

\[
\frac{\Phi_{\text{mis}}}{\Phi_{\text{mod}}} = \frac{(\tilde{N}_{H_2}O)_{\text{mis}}}{(\tilde{N}_{H_2}O)_{\text{mod}}}
\]

(16)

from which the columnar abundance \( (\tilde{N}_{H_2}O)_{\text{mis}} \) can be retrieved:

\[
(\tilde{N}_{H_2}O)_{\text{mis}} = (\tilde{N}_{H_2}O)_{\text{mod}} \frac{\Phi_{\text{mis}}}{\Phi_{\text{mod}}}
\]

(17)

4. Results and Data Processing

To test the performance of the proposed algorithm we attempted to estimate water vapor contents from AVIRIS data collected on 25 June 1987 over the Cuprite mining district in southwest Nevada [located at 37°45’ North (latitude) and 117°6’ West (longitude)]. Figure 6 shows an image (gray-scale) of the observed scene from which we can recognize some clouds (right, bottom) with their corresponding shadows.

In Fig. 7 a false-color water vapor columnar abundance map that we retrieved by applying this algorithm is shown. Note that the columnar water vapor value becomes small over clouds and highly variable over the clouds’ shadow. This behavior may be explained if we observe that clouds are atmospheric bod-
investigation because for optically thick clouds the absorption effect may be slightly enhanced owing to multiple scattering of solar radiation within the clouds. Note that the water vapor map of Fig. 7 contains some details of the original observed scene, an effect connected to the topography of the scene.

Figure 8(a) compares horizontal profiles of water vapor abundance (along the same line crossing the AVIRIS image) as retrieved from the two algorithms developed here. As can be seen, the two independent estimates agree fairly well. However, the water vapor columnar abundance estimates computed with the simple model are on average noisier and greater than those computed with the other method. This behavior is also confirmed from the corresponding histograms (empirical probability-density function) shown in Fig. 8(b). The curve that corresponds to the first method has a mean of 0.71 g cm$^{-2}$ and a standard deviation of 0.040 g cm$^{-2}$. The curve that corresponds to the simple method has a mean of 0.89 g cm$^{-2}$ and a standard deviation of 0.098 g cm$^{-2}$. In our belief this behavior originates from the too-coarse resolution of the AVIRIS, which samples the involved oxygen line with one spectral channel alone, hence making the line area estimation uncertain.

The tiny standard deviation of the computed H$_2$O abundance suggests that the retrieval algorithm is accurate and stable, and it also proves that the derived water vapor abundance is rather insensitive to changes of surface reflectance.

To better investigate the stability of the proposed algorithm with respect to target albedo variation, we have used the MODTRAN 4 code to simulate radiance spectra that correspond to different targets with constant albedo. The model utilized was the mid-latitude summer model, with 20 km of visibility and rural extinction.
Figure 9 shows the vertically integrated water vapor abundance retrieved from simulations, which is independent of the surface albedo. The maximum deviation from the expected horizontal straight line is less than 1%, a comfortable stability that was reached after subtraction of the radiation component backscattered by the atmosphere (path radiance) from the total at-sensor radiance. In this way the distortions of the line area that are due to large-angle scattering events were strongly reduced.

We have also tested the algorithm’s stability, utilizing simulated radiance from targets (e.g., a maple leaf spectrum) whose reflectance changes with wavelength. This test showed that the algorithm’s performance is not corrupted by inclusion of a spectrally selective target.

To comprehend the effect of atmospheric scattering on this kind of band measurement, we have plotted in Fig. 10 the ratio of the path radiance and the ground’s reflected radiance directly transmitted to the sensor.

Table 1. Main Characteristics of the VIRS-200 Imaging Spectrometer

<table>
<thead>
<tr>
<th>Characteristic</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Type</td>
<td>Push-broom</td>
</tr>
<tr>
<td>Number of channels</td>
<td>20 of 240 selectable</td>
</tr>
<tr>
<td>Free spectral range (μm)</td>
<td>0.4–1.0 (2.5-nm FWHM)</td>
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<td>Instantaneous field of view (mrad)</td>
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<td>Spatial sampling interval (mrad)</td>
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<td>Cross-track samples</td>
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<tr>
<td>Scan rate (scans/s)</td>
<td>12, 20, 30</td>
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<tr>
<td>Quantization accuracy</td>
<td>10 bits</td>
</tr>
<tr>
<td>Signal-to-noise ratio (at albedo 0.5)</td>
<td>20/400</td>
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</tbody>
</table>
versus the wavelength. As can be seen, this ratio increases toward line center owing to the augmented extinction coefficient that dims the farthest radiation sources \( /H_2O \) the ground reflected radiance while the nearest ones remain visible (backscattered contributions that originate near the sensor).

To test the empirical simple model for \( H_2O \) retrievals we also used a VIRS-200 imaging spectrometer to collect in-field data. The VIRS-200 is a push-broom imaging spectrometer equipped with a 2-dim array (CCD) of photosensitive elements and operating in the visible–near-infrared spectral range. The sensor digitizes 20 of the 240 available spectral channels, which are uniformly spaced from 400 to 1000 nm in 2.5-nm steps. The wavelengths of the 20 recorded channels, which are digitized within 10-bit accuracy, are freely chosen. The main sensor characteristics are detailed in Table 1, and additional information concerning this sensor is given elsewhere.\(^{37,38}\)

The selected spectral configuration for the VIRS-200 instrument was aimed to allow us to observe the \( O_2 \) absorption band located at 688 nm and the two \( H_2O \) absorption bands centered at 695 and 719 nm, as given in Table 2.

Measurements were performed in October 2003 at Florence, Italy, near 11:00 a.m. local time. We observed the sky’s diffuse radiation (downwelling path radiance) from ground level. Observations were completed with dark-signal measurements. Experimental data were averaged and corrected for the corresponding dark signal and the spectrally and spatially coherent noise pattern by means of a flat-field calibration procedure developed by Barducci and Pippi.\(^{39}\)

Figure 11 shows a spectrum of transmittance measured with VIRS-200 data. This spectrum was normalized to continuum introducing two linear interpolations, the first in the VIRS-200 channel from 684.75 to 712.25 nm and the second from 712.25 to 744.75 nm. The spectrum was compared with atmospheric transmittance simulated by the MODTRAN 4 radiative transfer code for a rural midlatitude winter atmosphere with 20 km of visibility and for a vertical path.

Applying the procedure described in Eq. (17), first we estimated \( \psi_{\text{mis}} \) and then we computed \( \psi_{\text{mod}} \) from the simulated spectrum to retrieve a columnar water vapor abundance \( (N_{H_2O})_{\text{mis}} \) of 2.10 g cm\(^{-2}\).

### Table 2. Spectral Configuration of the VIRS-200 Imaging Spectrometer

<table>
<thead>
<tr>
<th>Channel Number</th>
<th>Central Wavelength (nm)</th>
</tr>
</thead>
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<td>683.75</td>
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<tr>
<td>2</td>
<td>686.25</td>
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5. Concluding Remarks

The problem of estimating the abundance of important atmospheric constituents (e.g., water vapor) from remotely sensed hyperspectral images has been examined. A new spectroscopic method for the retrieval of atmospheric columnar water vapor at high spatial resolution was developed, and its performance and formulation were discussed. The algorithm compares the observed and expected (theoretical) line areas to calculate the integrated abundance of the absorber over the ray path traveled by the sensed radiation. To maintain the mathematical formulation of the procedure as simple as possible, we neglected the dependence of the cross section’s spectrum on atmospheric pressure and temperature (profile) for the absorption line. Early results obtained from processing of the water vapor absorption line at 940 nm as extracted from a hyperspectral image gathered by the AVIRIS were depicted, and they proved the effectiveness of the proposed algorithm.

We tested an additional empirical retrieval procedure with data gathered by the VIRS-200 imaging spectrometer to arrange for a faster although approximate estimate. This simpler retrieval algorithm does not require a priori knowledge of the cross-section spectrum of the absorption line to retrieve the columnar abundance of the related constituent. This method is based on the contemporary observation of two absorption lines, one of which, that originated from an atmospheric absorber at a fixed mixing ratio, is used as reference to compensate for most of effects of the viewing geometry and the total air mass traveled by the observed radiation. In our tests we chose as a reference the oxygen absorption line located near 760 nm.

Fair agreement between the two estimates was observed, even if the water vapor columnar abundance estimated with the simple empirical model was on average noisier and slightly more biased (greater) than that computed with the other algorithm. This behavior originated from the too coarse resolution of the AVIRIS, which sampled the O$_2$ line with one spectral channel alone, hence making the line-area estimation uncertain.

We also verified that the columnar water vapor retrieved with the new spectroscopic method was insensitive to variations in surface spectral reflectance, provided that scattering contributions to the at-sensor radiance are reliably corrected or were small enough that they do not greatly affect the observed line shape.

References

19. W. B. Grant, “Water vapor absorption coefficient in the