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1 Introduction

In this document we describe the ADV/ASV algorithm used for volcanic ash plume detection and retrieval of the plume properties. The algorithm uses data from visible (VIS) and near-infrared (NIR) channels of the Advanced Along Track Scanning Radiometer (AATSR) aboard ENVISAT for aerosol retrieval, and from the thermal infrared (TIR) channels data for ash plume detection. The aerosol optical depth (AOD) retrieval method, largely based on the dual view capability of AATSR, has been well established and reported in the scientific literature (Veefkind and de Leeuw, 1998a; Curier et al., 2009; Kolmonen et al., 2013a). Our emphasis in this work is on combining the ash detection using TIR channels with aerosol load information retrieved from the VIS/NIR channels, and in the special features related to volcanic aerosols.

The ash plume detection is primarily based on the brightness temperature difference (BTD) (Prata, 1989), but additional information from VIS/NIR channels is used as well. The aerosol optical depth (AOD) retrieval within the ash plume is conducted by the AATSR Dual View (ADV) and AATSR Single View (ASV) algorithms, originally designed at TNO (Veefkind and de Leeuw, 1998a) and further developed at FMI (Curier et al., 2009; Kolmonen et al., 2013a). After retrieving the AOD, information on the volcanic aerosol mass load can be derived, based on the aerosol models used in the retrieval and an assumption on the ash density. In addition to the plume position, we can acquire information on the plume height using the stereo view of AATSR. The height estimation algorithm is described in a separate document (Virtanen and de Leeuw, 2013).

Scope

In this document the theoretical basis of the aerosol optical depth (AOD) retrieval algorithm based on the AATSR visible (VIS) and near-infrared (NIR) channels is described. The AATSR dual view (ADV) and AATSR single view (ASV) algorithms are originally designed for retrievals of anthropogenic/natural aerosols of different nature (Veefkind and de Leeuw, 1998a; Curier et al., 2009; Kolmonen et al., 2013a), and the necessary modifications for retrievals of volcanic ash specific aerosols are described.

The conversion of the AOD to volcanic ash mass load based on the aerosol models employed is explained. The ash detection method, primarily based on the brightness temperature difference (BTD), is described, with discussion of false alerts. A brief description of the AATSR instrument is given, and an account of the product output format is given.

The algorithm is still under development, and this first version of the ATBD will be updated as needed. Parts where improvement is needed are discussed in the text. Examples of volcanic ash plumes are shown in Fig. 1.

Figure 1: May 6th 2010, Eyjafjallajökull eruption; October 28th 2002, Etna eruption; May 5th 2008, Chaiten eruption.
2 AATSR instrument

AATSR (Advanced Along Track Scanning Radiometer) on board the ENVISAT satellite (2002-2012), was designed to measure sea surface temperature (SST) with accuracy higher than 0.3 K for use in studies of climate change (ESA, 2007). The AATSR instrument has seven channels centered at wavelengths of 0.555, 0.659, 0.865, 1.61, 3.7, 10.85, and 12.0 \( \mu \text{m} \) (Table 1). It uses a conical scanning geometry, which means that the same pixel on the ground is viewed twice, once with a \( \sim 55^\circ \) forward view, and about two minutes later by a near-nadir view. The actual viewing zenith angle (VZA) \( \mu \) depends on the across track position for both forward and near-nadir views. The AATSR swath width is 512 km and it has a revisit time of three days (at mid-latitudes), resulting in global coverage in 5-6 days. Together with its predecessor ATSR-2 on ERS-2 (1995-2011) it spans a time period of 17 years (1995-2012).

<table>
<thead>
<tr>
<th>Channel</th>
<th>Centre Wavelength</th>
<th>Bandwidth</th>
<th>Performance</th>
<th>Application</th>
</tr>
</thead>
<tbody>
<tr>
<td>VIS</td>
<td>0.55 ( \mu \text{m} )</td>
<td>0.555 ( \mu \text{m} )</td>
<td>20 nm</td>
<td>SNR &gt; 20</td>
</tr>
<tr>
<td></td>
<td>0.66 ( \mu \text{m} )</td>
<td>0.659 ( \mu \text{m} )</td>
<td>20 nm</td>
<td>SNR &gt; 20</td>
</tr>
<tr>
<td></td>
<td>0.87 ( \mu \text{m} )</td>
<td>0.865 ( \mu \text{m} )</td>
<td>20 nm</td>
<td>SNR &gt; 20</td>
</tr>
<tr>
<td>NIR</td>
<td>1.6 ( \mu \text{m} )</td>
<td>1.61 ( \mu \text{m} )</td>
<td>0.3 ( \mu \text{m} )</td>
<td>SNR &gt; 20</td>
</tr>
<tr>
<td>SWIR</td>
<td>3.7 ( \mu \text{m} )</td>
<td>3.70 ( \mu \text{m} )</td>
<td>0.3 ( \mu \text{m} )</td>
<td>NE( \Delta T ) &lt; 80 mK</td>
</tr>
<tr>
<td></td>
<td>11 ( \mu \text{m} )</td>
<td>10.85 ( \mu \text{m} )</td>
<td>1.0 ( \mu \text{m} )</td>
<td>NE( \Delta T ) &lt; 50 mK</td>
</tr>
<tr>
<td>TIR</td>
<td>12 ( \mu \text{m} )</td>
<td>12.00 ( \mu \text{m} )</td>
<td>1.0 ( \mu \text{m} )</td>
<td>NE( \Delta T ) &lt; 50 mK</td>
</tr>
</tbody>
</table>

Table 1: AATSR spectral channels, and their use in this work.

Although the instrument was designed primarily for SST studies, the visible bands allow for its use in vegetation index studies. It is also successfully used for atmospheric aerosol retrieval (Flowerdew and Haigh, 1995), (Voelpkind and de Leeuw, 1998a), (Grey et al., 2006). The basic principle of the retrieval is to match (A)ATSR top of atmosphere (TOA) reflectance, derived from the measured radiance, to modeled reflectance. The modeled reflectance is computed applying a radiation transfer code for the propagation of the solar radiance through the atmosphere. Cloud screening is needed as the aerosol properties can be retrieved only for cloud-free sky.

In this work, we focus on the retrieval of volcanic ash. The first four channels are used to provide the ratio of (upward) reflected radiation to the incoming solar radiation at the top of atmosphere, i.e. the TOA reflectance \( R \), while the latter three channels provide information of the surface temperature via brightness temperatures \( T \). The reflectance (visible) channels are used for the retrieval of aerosol properties using the ADV algorithm. The use of TIR channels is limited to the detection of the ash plumes (and plume top height estimates), but with similar techniques they could be used for AOD retrieval as well. The visible channels can be used for aerosol retrievals only on day-time orbits, while the thermal channels can be used for ash plume detection also on night time orbits.

2.1 SLSTR

The connection with ENVISAT was lost in April 2012, so AATSR data is not available for monitoring future volcanic activity, and can only be used to study previous eruptions. However, its successor, the Sea and Land Surface Temperature Radiometer (SLSTR) is scheduled for launch in 2014 onboard Sentinel-3. The SLSTR carries the heritage from the ATSR series with similar characteristics, but it also has many improvements (Donlon et al., 2012).

- The swath will be wider, \( \sim 1400 \text{ km} \) for nadir view and for \( \sim 740 \text{ km} \) for dual view, compared to 500 km of AATSR.
- The resolution is increased to 0.5 km for the visible channels, compared to 1 km resolution of AATSR.
- There will be more channels. Two new channels in the short wave infrared (SWIR) are centered at 1.375 \( \mu \text{m} \) and 2.25 \( \mu \text{m} \). In addition two dedicated fire channels are included.
- Two instruments. The plan is to have two instruments in orbit aboard Sentinel-3a and Sentinel-3b, respectively.

Preparation for the use of SLSTR for ash retrieval is in progress.
3 ADV/ASV algorithm

The ADV and ASV algorithms for global aerosol optical depth retrieval were originally developed by Veefkind et al. (Veefkind and de Leeuw, 1998a,b), and has been extensively developed by the FMI/University of Helsinki group in recent years (Kolmonen et al., 2013a). Recent improvements include implementation of online mixing of four different aerosol types in the retrieval, per-pixel error characterization, enhanced post-processing cloud screening, and generation of several new output products, e.g. surface reflectance and single scattering albedo. Several minor corrections have also been made. The present AOD product is extensively validated in the ESA Aerosol-CCI project (de Leeuw et al., 2013; Holzer-Popp et al., 2013).

The algorithm consists of two parts: AATSR dual view (ADV) algorithm, which is used over land, and AATSR single view algorithm (ASV) which is used over ocean. The reason for this duality is that while the ocean surface can be reasonably well modeled allowing for AOD retrieval using single view only, the same is not true for various land surfaces. The key feature of the AATSR instrument in AOD retrievals is its stereo view capability, which allows us to eliminate the surface reflectance from the retrieval equations, using certain assumptions.

In the following two sections (3.1 and 3.2) we will describe the land and ocean retrieval algorithms. In section 4 we will discuss how the standard algorithm needs to be modified when concentrating on volcanic ash. We will also discuss the limitations of visible wavelength AOD retrievals using the ADV/ASV algorithm.

3.1 ADV algorithm for over land retrieval

This section is largely based on the ESA-Aerosol-CCI document by (Kolmonen et al., 2013b), and to the recent paper by (Kolmonen et al., 2013a).

3.1.1 Formal background of the dual-view algorithm

The dual-view algorithm for the AATSR top-of-atmosphere (TOA) reflectance is meant for retrieval of aerosol optical properties over land (Veefkind and de Leeuw, 1998a; Veefkind et al., 2000; González, 2003; Curier et al., 2009; Kolmonen et al., 2013a). These properties include AOD for three wavelengths (nominally 0.555, 0.659 and 1.61 µm). In addition, an aerosol model is retrieved. The aerosol model used in retrievals specific for volcanic ash is described later in section 4. Here we describe the general model used in retrievals of anthropogenic and natural aerosols.

The general aerosol model is a mixture of four aerosol components, each of which is described by a lognormal size distribution defined by an effective radius and standard deviation (see section 3.1.2), and a complex refractive
Two of the aerosol components define small particles and the other two coarse particles. One of the small particle components is non-absorbing while the other one is strongly absorbing. By mixing these two components, arbitrary absorbing properties for the small particles can be set. The coarse particle components are sea salt and dust. Of the four components the dust one is composed of non-spherical particles, which is important for the computation of the optical properties. The final aerosol model is determined by first mixing the small and large components respectively, and finally mixing the ensuing small and coarse models together.

Figure 3: The ADV/ASV aerosol model generally used in global retrievals is a mixture of four components: 1) non-absorbing fine, 2) absorbing fine, 3) sea salt, and 4) dust. The first two compose the fine mode particles while the latter two make up the coarse mode particles. The mixture between the fine and coarse components is defined by the fine mode fraction $b_1$. The internal mixing of the fine mode components is determined by parameter $b_2$, while the internal mixing between the two coarse mode components is governed by the dust fraction $b_{	ext{dust}}$ (which is not retrieved but taken from climatology).

The aerosol components are adopted from the ESA Aerosol CCI (Climate Change Initiative) project. The properties of the components are described in Table 2. One of the needed mixtures, the dust fraction, is taken from the AEROCOM aerosol climatology and Aeronet (Holzer-Popp et al., 2013). The non-absorbing fine - absorbing fine mixture is retrieved semi-freely. The mixture can have any value in the range of $-0.3$ from the AEROCOM climatology value. The fine - coarse mixture is retrieved completely independent of the climatology. The algorithm is based on a number of assumptions:

- **TOA reflectance $\rho$** is of the form (Veefkind et al., 2000)
  \[
  \rho(\mu_1, \mu, \phi, \lambda) = \rho_a(\mu_1, \mu, \phi, \lambda) + \frac{T(\mu_1, \mu, \phi, \lambda)\rho_a(\mu_1, \mu, \phi, \lambda)}{1 + \tau(\lambda)R_a(\lambda)},
  \]
  where $\rho_a$ is the reflectance due to the atmosphere, $\rho_g$ is the ground reflectance, $T$ is the multiple of downward and upward atmospheric transmittance, $s$ is the atmospheric backscatter ratio, and $R_a$ is the surface albedo. Reflectance and transmittance parameters: $\mu_1$ is the solar zenith angle, $\mu$ is the viewing (satellite) zenith angle, $\phi$ is the relative azimuth angle between sun and the satellite, and $\lambda$ is the wavelength. Note that multiple scattering between ground and atmosphere is assumed here to be angle independent for method development purposes. It has also been suggested that the multiple scattering in the surface-atmosphere system will lead to isotropically distributed scattering (Wanner et al., 1997).

- **Atmospheric reflectance**
  \[
  \rho_a(\mu_1, \mu, \phi, \lambda) = \rho_R(\mu_1, \mu, \phi, \lambda) + \rho_{\text{aer}}(\mu_1, \mu, \phi, \lambda),
  \]
  where $\rho_R$ is reflectance due to Rayleigh scattering and $\rho_{\text{aer}}$ is reflectance due to aerosols.

- **Reflectance due to aerosols** is computed using the modified linear mixing method by (Abdou et al., 1997). The method as adapted to ADV for two aerosol components is
  \[
  \rho_{\text{aer}} = b_1 \frac{\omega_1}{\omega_1} e^{-\tau_1 |\omega_1 - \omega_\text{mix}|} \rho_1 + b_2 \frac{\omega_2}{\omega_2} e^{-\tau_2 |\omega_2 - \omega_\text{mix}|} \rho_2,
  \]
  where $\omega$ is the single scattering albedo (SSA) and $\tau$ is AOD. Subscripts 1 and 2 refer to the aerosol types while mix refers to the linear mixture of the two types, $\omega_\text{mix} = b_1 \omega_1 + b_2 \omega_2$. For the weighting coefficients $b_1 + b_2 = 1$. The modified linear mixing method is applied to take better into account the effects of mixing two aerosols with different absorbing properties. This is done by introducing the single scattering albedo into linear mixing. If the SSAs of the aerosol types are identical, equation (3) simplifies to
  \[
  \rho_{\text{aer}} = b_1 \rho_1 + b_2 \rho_2.
  \]

The needed aerosol transmittance is computed using linear mixing.
The ratio $k$ between the ground reflectance of the forward and nadir views is independent of wavelength (Flowerdew and Haigh, 1995):

$$k = \frac{\rho_f(\mu_1, \mu, \phi, \lambda)}{\rho_n(\mu_1, \mu, \phi, \lambda)}$$  \hspace{0.5cm} (5)

where $\rho_f$ and $\rho_n$ are the forward and nadir ground reflectance, respectively. Also, because reflectance due to aerosols at 1.61 $\mu$m is small compared to ground reflectance, the $k$-ratio is computed using equation

$$k = \frac{\rho_f(\mu_1, \mu, \phi, 1.61 \mu m)}{\rho_n(\mu_1, \mu, \phi, 1.61 \mu m)}$$  \hspace{0.5cm} (6)

Coarse particles contribute to reflectance at IR wavelengths. This may cause some error in the $k$-ratio for instance in heavy dust conditions or with volcanic ash plumes.

The 0.865 $\mu$m channel is excluded from the retrieval because the $k$-ratio assumption is usually not valid as there is a strong reflectance by vegetation at this wavelength (González et al., 2000).

The dual-view method for AOD retrieval is derived based on the above assumptions. Equation (1) can be written separately for the forward and nadir views. Then, by combining these equations while keeping in mind that the multiple scattering is assumed to be angle independent, relation

$$\frac{\rho_n(\mu_1, \mu, \phi, \lambda) - \rho_n(\mu_1, \mu, \phi, \lambda)}{\rho_n(\mu_1, \mu, \phi, \lambda)T_n(\mu_1, \mu, \phi, \lambda)} = \frac{\rho_f(\mu_1, \mu, \phi, \lambda) - \rho_f(\mu_1, \mu, \phi, \lambda)}{\rho_f(\mu_1, \mu, \phi, \lambda)T_f(\mu_1, \mu, \phi, \lambda)}$$  \hspace{0.5cm} (7)

can be made formally. The key aspect of the dual-view algorithm is to introduce the $k$-ratio in equation (7) to obtain

$$\frac{\rho_n(\mu_1, \mu, \phi, \lambda) - \rho_n(\mu_1, \mu, \phi, \lambda)}{T_n(\mu_1, \mu, \phi, \lambda)} = \frac{\rho_f(\mu_1, \mu, \phi, \lambda) - \rho_f(\mu_1, \mu, \phi, \lambda)}{kT_f(\mu_1, \mu, \phi, \lambda)}.$$  \hspace{0.5cm} (8)

Now all the needed knowledge about ground reflectance is in the $k$-ratio.

### 3.1.2 Computational aspects of ADV

Modeled values of the atmospheric reflectance $\rho_a$ and transmittance $T$ must be determined in order to use equation (8) for the retrieval of aerosol properties. These values, together with other information can be computed using radiative transfer (RT) methods. RT methods provide a way to solve the forward problem of the retrieval. The forward problem for the case of the atmospheric reflectance can be stated as: when the atmospheric conditions (aerosol and gas concentrations) are known, determine the amount of light that is reflected from the atmosphere towards a satellite instrument.
During a retrieval the forward problem must be solved numerous times, which is time consuming. The most common technique to overcome this is to perform the radiative transfer calculations for a set of fixed variables before the retrieval. The calculated values are arranged into an array that is called a look-up-table (LUT). During the retrieval the needed values can then be interpolated quickly from the LUT. The chosen RT algorithm that is used with the ADV is the DAK (Doubling Adding KNMI, de Haan et al. 1987).

<table>
<thead>
<tr>
<th>model</th>
<th>( r_g ) ((\mu m))</th>
<th>( \sigma )</th>
<th>( m )</th>
<th>ALH (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>non-absorbing fine</td>
<td>0.07</td>
<td>1.700</td>
<td>1.40 - 0.003i</td>
<td>0-2</td>
</tr>
<tr>
<td>absorbing fine</td>
<td>0.07</td>
<td>1.700</td>
<td>1.50 - 0.040i</td>
<td>0-2</td>
</tr>
<tr>
<td>sea salt</td>
<td>0.788</td>
<td>1.822</td>
<td>1.40 - 0.000i</td>
<td>0-1</td>
</tr>
<tr>
<td>dust</td>
<td>0.788</td>
<td>1.822</td>
<td>1.56 - 0.002i</td>
<td>2-4</td>
</tr>
</tbody>
</table>

Table 2: The Aerosol CCI aerosol components (de Leeuw et al., 2013). Listed are the geometric radius \( r_g \), standard deviation \( \sigma \), refractive index \( m \), and the aerosol layer height (ALH). For ash-specific retrievals alternate aerosol models are used, see section 4.

LUTs are computed for each aerosol component. The size distribution of an aerosol component is described by a log-normal number size distribution of the form

\[
\frac{dN}{d\ln r} = \frac{N_0}{\ln \sigma \sqrt{2\pi}} \exp \left( -\frac{\ln^2(r/r_g)}{2\ln^2 \sigma} \right),
\]

where \( r \) is the particle radius. The size distribution of an aerosol type is defined by the geometric mean radius \( r_g \) and standard deviation \( \sigma \) (Heitzenberg, 1994). The total number of aerosol particles \( N_0 \) depends on the aerosol load. Aerosol optical properties are computed by applying Mie calculations (Mie, 1908) except for the non-spherical dust component where T-matrix was used (Mischenko and Davis, 1994). These calculations require the knowledge of the aerosol particle size distribution and refractive index. The parameters describing the four aerosol components are provided in Table 2.

The LUTs are computed for discrete sun zenith, viewing zenith and relative azimuth angles, for each AATSR VIS/NIR wavelength, and for a number of reference AOD levels. Currently for ADV 15 discrete values ranging from 0 to 90° for zenith angles and 19 discrete values between 0 and 180° for the azimuth angle are used. Typically ten AOD levels ranging from 0.05 to 4.0 at \( \lambda = 0.500 \mu m \) are used with the ADV. Transmittance and reflectance are computed also for Rayleigh (gas) scattering in standard atmospheric conditions. The maximum values of the sun and viewing zenith angles are currently limited to 75° used in the radiative transfer computations, to avoid effects due to refraction.

Equation (8) shows that the computational task is to find the aerosol type mixture and reference AOD level that solve the equation for all three\(^2\) AATSR wavelengths simultaneously. Due to measurement and modeling errors this task is impossible in practice. Instead, the task can be converted to a least-squares type of problem

\[
\arg_{b_1,b_2,L} = \min \sum_{i=1}^{N} \left[ \frac{\rho^g(\lambda_i) - \rho^g_0(b_1,b_2,L,\lambda_i)}{T_n(b_1,b_2,L,\lambda_i)} - \frac{\rho^f(\lambda_i) - \rho^f_0(b_1,b_2,L,\lambda_i)}{kT_n(b_1,b_2,L,\lambda_i)} \right]^2,
\]

where the fraction of the fine mode particles is \( b_1 \in (0,1) \), the non-absorbing component in fine particle mixture is \( b_2 = b_{2,A} \pm 0.3 \) with \( b_2 \in (0,1) \), and the reference AOD level is \( L \in (1,10) \). The mixture \( b_{2,A} \) is the AEROCOM a priori value. Note that the dust fraction is not retrieved but the AEROCOM climatology value is used. The angle arguments \((\mu_1,\mu,\phi)\) have been omitted for brevity. The number of wavelengths \( N_\lambda = 3 \). Equation (10) also shows that the modeled atmospheric reflectance and transmittance are now functions of the decision arguments \( b_1, b_2 \) for aerosol component mixtures and \( L \) for the reference AOD level. The task is now to find the decision arguments \((b_1,b_2,L)\) that minimize the least-squares sum.

With each mixture, the modified linear mixing method that was introduced in equation (3), is applied for the computation of reflectance.

The minimization problem (10) can be optimized by applying mathematical optimization methods. Here the chosen method is Levenberg-Marquardt (see for example Gill et al. 1999). It is a trust-region type method well suited for least-squares problems, and is meant for unconstrained optimization. The latter feature causes additional considerations as the decision arguments are all box-constrained. This is handled in the evaluation of the least-squares sum where strict barrier functions are used for the constraining [7].

Another feature of the Levenberg-Marquardt method is that it is a local optimizer. It will converge efficiently to the nearest local minimum. To increase the probability of finding the globally best solution an initial search is done in a limited discrete set of decision parameters: ten mixtures \( b_1, b_2 \), and ten AOD levels \( L \). The results of the search are then used as the initial guess for the Levenberg-Marquardt method.

When all the decision parameters are set during the retrieval, the resulting AOD \( \tau \) can be computed from the LUTs

\[
\tau(\lambda) = b_1[b_2\tau_{naf}(\lambda,L) + (1-b_2)\tau_{df}(\lambda,L)] + (1-b_1)[b_{dust}\tau_{dust}(\lambda,L) + (1-b_{dust})\tau_{ss}(\lambda,L)],
\]

\(^2\)As mentioned before, the AOD over land is retrieved for only wavelengths 555 nm, 659 nm and 1.61 \(\mu m\).
where the abbreviations are: naf - non-absorbing fine component, af - absorbing fine component, and ss - sea salt coarse component. Dust fraction $b_{\text{dust}}$ is the above mentioned AEROCOM a priori value.

### 3.1.3 Error estimation for ADV

In this section the effect of AATSR measurement error on AOD is described. First, the uncertainty for the retrieved aerosol model decision parameters, which include the fine mode fraction $b_1$, the absorbing/non-absorbing fine particle mixture $b_2$, and the AOD level $L$, is determined. Then these errors are used to determine the uncertainty in the retrieved AOD.

The other possible sources for errors arise from modeling. These include uncertainty in the aerosol model selection (fine mode fraction, absorbing/non-absorbing fine particle mixture, dust fraction), LUT interpolation errors, and radiative transfer computation errors. Formal treatment is based on the general equation formalism by Tarantola (1987) (pp. 77 - 82). First, denote the parameters in the least squares problem (10) by

$$ x = \{b_1, b_2, L, r\}, $$

and the problem equations by

$$ f_i(x) = \frac{\rho_i(\lambda_i) - \rho_i^b(b_1, b_2, L, \lambda_i)}{T^n(b_1, b_2, L, \lambda_i)} - \frac{\rho_i(\lambda_i) - \rho_i^a(b_1, b_2, L, \lambda_i)}{kT^t(b_1, b_2, L, \lambda_i)} $$

where $r = \{\rho^n(\lambda_i), \rho^a(\lambda_i)\} \forall i \in (1.3)$ is the measured nadir and forward reflectance, $b_1$ is the fine mode fraction, $b_2$ is the absorbing/non-absorbing fine particle mixture, and $L$ is the LUT AOD level. Index $i$ refers to the wavelengths; $i = 1, 2, 3$. Note that the dust fraction is not retrieved, it is taken from the AEROCOM climatology. This kind of formulation of the problem enables the determination of uncertainty for the decision parameters based on the measurement error. The formulation could take into account the effect of a priori information for $b_1, b_2$ and $L$, but this is neglected as the only error is assumed to come from the measurement.

Equation (13) can be solved in least-squares sense using a quasi-Newton method. The maximum likelihood point can be found using iteration

$$ x_{n+1} = C_X F_n^T (C_T + F_n C_X F_n^T)^{-1} f(x_n), $$

where

$$ F_n = \left( \frac{\partial f}{\partial x} \right)_{x_n}. $$

A posteriori covariance is

$$ C_{X'} = (F_{\infty} C_T^{-1} F_{\infty} + C_X^{-1})^{-1}, $$

where $x_{\infty}$ is the solution of the minimized equation (13). Note that, even though the ADV solution is not computed, using the iteration scheme above to determine the a posteriori covariance is still possible.

The Jacobian matrix $F$ is of the form

$$ F = \begin{pmatrix} \frac{\partial f_1}{\partial x_1} & \frac{\partial f_1}{\partial x_2} & \frac{\partial f_1}{\partial L} \\ \frac{\partial f_2}{\partial x_1} & \frac{\partial f_2}{\partial x_2} & \frac{\partial f_2}{\partial L} \\ \frac{\partial f_3}{\partial x_1} & \frac{\partial f_3}{\partial x_2} & \frac{\partial f_3}{\partial L} \end{pmatrix}. $$

All the partial derivates are computed numerically as the evaluation of these values requires interpolation from the aerosol LUTs and analytical differentiation is impossible.

The covariance $C_T$ here consists of only measurement errors. For AATSR this error is taken to be 5% of the measured signal for the whole spectrum (Thomas et al., 2009). The principal difficulty is that in equation (13) there are two measured values $\rho^n(\lambda_i)$ and $\rho^a(\lambda_i)$. The formulation in equation (16), however, takes into account the uncertainty of only one value in the covariance matrix. The solution is that the larger absolute value of the nadir and forward relative measurement errors is used. It would be useful to study individually the effect of the nadir and forward measurement error on the aerosol model parameters in future. In addition, when all errors are considered to be Gaussian in nature, modeling errors could be simply added to the measurement errors (Tarantola, 1987). Another simplification made here is that measurement errors do not correlate. Thus, $C_T$ is diagonal. This assumption does not hold true for the a posteriori covariance $C_{X'}$. The uncertainty in the aerosol model parameters will be correlated.

The a priori covariance matrix for the aerosol model defining parameters $C_X$ is neglected at the moment as the uncertainty contribution of the measurement error to these very parameters is the motivation of this exercise.

AOD is determined for each of the three wavelengths using the aerosol model defined by the three optimized decision parameters: $b_1$, $b_2$ and $L$. First, for all four aerosol types that are used, the corresponding AOD is interpolated from their LUTs by using the AOD level parameter $L$. Then, simultaneously, fine aerosol type is mixed from the non-absorbing and absorbing AOD using $b_2$, and coarse aerosol type is mixed from the sea salt.
and dust AOD using the dust fraction. The final AOD is then mixed from the fine and coarse AOD using $b_1$; see equation (11).

Denote the AOD interpolation/mixing operator by $p$. Then for wavelength $i$, the AOD is

$$
\text{AOD}_i = p_i(b_1, b_2, L).
$$

(18)

The covariance of AOD is then

$$
\text{C}_{\text{AOD}} = p \text{C}_X p^T,
$$

(19)

where $p$ is now the Jacobian of the interpolation/mixing operator:

$$
p = \begin{pmatrix}
\frac{\partial p_1}{\partial b_1} & \frac{\partial p_1}{\partial b_2} & \frac{\partial p_1}{\partial L} \\
\frac{\partial p_2}{\partial b_1} & \frac{\partial p_2}{\partial b_2} & \frac{\partial p_2}{\partial L} \\
\frac{\partial p_3}{\partial b_1} & \frac{\partial p_3}{\partial b_2} & \frac{\partial p_3}{\partial L}
\end{pmatrix}
$$

(20)

For reference, see e.g. (Meyer, 1975). The uncertainty estimate for AOD can be finally found in the diagonal of $\text{C}_{\text{AOD}}$.

The uncertainty of the dust fraction, which is not a retrieved parameter, could be added to $\text{C}_X$, if this uncertainty was assumed to be Gaussian.

### 3.2 ASV algorithm for over ocean retrieval

The basic principle of the algorithm is to minimize the discrepancy between the TOA measured and modeled reflectances at wavelengths of 0.555, 0.659, 0.865 and 1.61 $\mu$m. The modeled reflectance is described below.

The aerosol modeling follows the description given for the ADV algorithm, i.e. the aerosol model is based on the three mixtures that are introduced with equation (11).

#### 3.2.1 The modelled TOA reflectance

The TOA reflectance over ocean is given by (Veefkind and de Leeuw, 1998a):

$$
\rho_{\text{TOA}} = \rho_a + T_{\downarrow} \rho_{s,\text{dir}}(1 - S \rho_{s,\text{dir}}) T_{\uparrow} + t_{\downarrow} \rho_{s,\text{dif}} T_{\uparrow} + T_{\downarrow} \rho_{s,\text{dif}} t_{\uparrow} + t_{\downarrow} \rho_{s,\text{iso}} t_{\uparrow},
$$

(21)

where $\rho_{\text{TOA}}$ is the top-of-the atmosphere reflectance, $S$ is the spherical albedo, $T$ is the direct transmittance and $t$ is the diffuse transmittance upwards ($\uparrow$) and downwards ($\downarrow$). The terms $\rho_a$ and $\rho_s$ are the atmospheric and surface reflectances, respectively, and the other terms come from the ocean surface model which is described in the next section. The multiple scattering between surface and atmosphere has been included only for the direct down direct up case as it becomes negligible when diffuse transmittance is applied. Note that geometric and wavelength dependencies in equation (21) are omitted for brevity. Explanation of the components in equation (21) in order from left to right are

- Reflectance due to scattering in the atmosphere by aerosols and molecules.
- Photons transmitted downward, reflected by the ocean surface, and transmitted up.
- Photons scattered along the downward path, reflected by the ocean surface, and transmitted up.
- Photons transmitted downward, reflected by the ocean surface, and scattered towards the satellite instrument.
- Photons scattered along the downward path, reflected by the ocean surface, and scattered towards the satellite instrument.

Each of the terms in equation (21) contains contributions of specular (Fresnel) reflection, oceanic whitecaps and subsurface scattering.

#### 3.2.2 Ocean reflectance modelling

The ocean surface reflectance is modeled as the sum of specular (Fresnel) reflectance (Cox and Munk, 1954) and reflectance by subsurface scattering. The Fresnel part is described by the geometric situation while the subsurface scattering is a function of chlorophyll concentration. The surface reflectance is a sum of four components based on atmospheric transmittance, see Eq. (21). The reflectance in these components is given by:

$$
\rho_{s,\text{dir}}(\mu_0, \mu, \phi, \lambda) = \rho_{\text{glint}}(\mu_0, \mu, \phi, \lambda) + \rho_{\text{chl}}(C, \lambda),
$$

(22)

where $\rho_{\text{glint}}$ is the sun glint and $\rho_{\text{chl}}$ is the subsurface reflectance due to chlorophyll concentration $C$, and it is assumed here to be Lambertian (Veefkind and de Leeuw, 1998a). In practice the reflectance due to sun glint is not taken into account because pixels flagged as sun glint in the AATSR L1 data are not used in the retrieval.
The geometric situation is described by the cosine of the solar zenith angle $\mu$, the cosine of the viewing zenith angle $\mu$, and the relative azimuth angle $\phi$. Reflectance depends on the wavelength $\lambda$. Subsurface reflectance is modeled after (Morel, 1988) for case I waters as

$$\rho_{s,\text{diff}}(\mu_0, \mu, \phi, \lambda) = \rho_{\text{Fresnel}}(\mu_0) + \rho_{\text{chl}}(C, \lambda)$$  \hspace{1cm} (23)

$$\rho_{s,\text{diff}}(\mu_0, \mu, \phi, \lambda) = \rho_{\text{Fresnel}}(\mu_0) + \rho_{\text{chl}}(C, \lambda)$$  \hspace{1cm} (24)

$$\rho_{s,\text{iso}}(\mu_0, \mu, \phi, \lambda) = 0.066 + \rho_{\text{chl}}(C, \lambda)$$  \hspace{1cm} (25)

In these equations $\rho_{\text{Fresnel}}$ is the Fresnel reflectance, and the factor 0.066 has been adapted from (Ivanov, 1975). The possible error caused by the approximate value is minimal because the contribution of the last term to the TOA reflectance is small. All of the above components include the contribution of the whitecap reflectance determined by the fraction of the ocean surface covered by whitecaps. The whitecap fraction $W$ is a function of wind speed $U$ (Monahan and O’Muircheartaigh, 1980):

$$W = 3.84 \times 10^{-6} \times U^{3.41}.$$

### 3.2.3 ASV in practice

In the ASV retrieval the same aerosol look-up-tables are used as for the ADV retrieval. Note, that there is no distinction between land and ocean retrieval with respect to aerosol components. AEROCOM a priori values and retrieval itself decide the aerosol composition for any given pixel.

As was mentioned above, the ASV method is based on minimizing the TOA measured and modeled reflectances. This leads to a minimization scheme, which is considerably different from that for ADV, which can be seen in equation (10). The main physical difference is that only one of the AATSR views is used. Currently the forward view is employed as it is less hindered by sun glint than the nadir view. The minimization in the ASV problem, following the ADV notation, is

$$\arg_{b_1, b_2, L} \min_{i=1}^{N_A} \sum_{i=1}^{N_A} \left[ \rho_i^\prime(\lambda_i) - \rho_{\text{TOA}}^\prime(b_1, b_2, L, \lambda_i) \right]^2,$$  \hspace{1cm} (27)

with the modifications that $N_A = 4$, as the 865 nm wavelength is also used, and $\rho_{\text{TOA}}^\prime$ TOA from equation (21) is now the combined atmospheric and ocean surface reflectance.

### 3.2.4 Error estimation for ASV

The effect of AATSR measurement error on the retrieved AOD was described for the ADV algorithm in section 3.1.3. This error treatment can be straightforwardly applied for the ASV by replacing equation (13) with the ASV minimization from equation (27):

$$f_i(x) = \rho_i^\prime(\lambda_i) - \rho_{\text{TOA}}^\prime(b_1, b_2, L, \lambda_i).$$  \hspace{1cm} (28)

### 3.3 Cloud Screening

Clouded pixels have to be excluded from retrieval as they mask the other contributions from the atmosphere to the measured TOA reflectance. The tests that are described here were designed for the use with ATSR-2 data. For AATSR cloud flags are included in the reflectance data (ESA, 2007). These flags were found to be too restricting: a significant amount of pixels that otherwise gave good validation results were excluded. The use of these flags will be studied more in future.

Presently, three separate cloud screening tests are used. These tests are based on the work of Saunders et al. (1988) and Koelmeijer et al. (2001). To automate the cloud screening, AATSR orbits are divided into scenes of 512×512 pixels. Reflectance in each of the scenes is histogrammed and thresholds or rejection values for the tests are determined from the histograms. The automation of the tests is described by González (2003). Brief description of the tests:

1. The gross cloud test. At the AATSR 12 $\mu$m brightness temperature channel clouds appear cooler than the underlying surface during day time. If the brightness temperature for a pixel is below threshold, the pixel is flagged as cloudy.

2. Generally, clouds are brighter than the underlying surface. If the reflectance of the 0.659 $\mu$m channel for a pixel is higher than threshold, the pixel is flagged as cloudy.

3. Ratio of the 0.865 and 0.659 $\mu$m reflectance. If the ratio is around one for a pixel, the pixel is flagged as cloudy. The distance from unity that governs cloud flagging is determined by the automation.

These tests are applied for both AATSR views. If any of the tests indicates that a pixel is clouded, it will be excluded from the retrieval. Note that for the ash specific retrievals the cloud test described above cannot be used, since they tend to misidentify ash plumes as clouds.
3.3.1 Additional post processing cloud screening

For each 10 × 10 km² superpixel retrieved with ADV a cloud post-processing test is applied to determine and discard the pixels that might potentially include cloud edge. Each pixel retrieved is analyzed together with the eight surrounding pixels. If, in addition to the tested pixel, less than 3 pixels are retrieved in the area, the tested pixel is considered to be a cloud edge and discarded. If, besides the tested pixel, at least 3 more pixels are retrieved and AOD for the tested pixel is smaller than 0.5, the tested pixel passes the cloud processing test. If AOD > 0.5, an additional standard deviation test is applied. If the standard deviation of AOD in the area is larger than 0.25, the tested pixel is discarded. These numbers are a compromise between global coverage and acceptable validation results.

3.4 Averaging of measured reflectance for ADV and ASV

measured TOA reflectance over the superpixel are described. Also the choice of aerosol models is discussed.

The retrieval process described above is highly time consuming, mainly due to optimization of three different parameters. For larger datasets, the use of a larger result pixel (superpixel) is necessary. In this section the methods for averaging the AATSR measured TOA reflectance over the 0.1° × 0.1° superpixel are described. In the ash-specific retrievals the features of the ash plumes are often so small that the full resolution is needed (fine resolution mode). On the other hand, the ash related datasets are limited in size, so that the processing times remain reasonable. However, some of the statistical methods used in the superpixel-averaging could be used in the ash retrievals for quality assurance purposes.

The natural assumption when averaging the TOA measured reflectance is that reflectance due to the atmosphere is sufficiently uniform over the averaged area. Here the term sufficient describes situations where sharp spatial gradients in aerosol conditions inside the area are not present. Reflectance due to atmospheric gases is assumed to be constant.

For the surface reflectance, however, this assumption can generally not be made. The complications in the averaging of the measured TOA reflectance are caused by the k-ratio approach of the ADV. The k-ratio is determined by applying equation (6) and using the nadir and forward view ground reflectance at 1.61 µm. It would be unrealistic to assume that ground reflectance is constant over the larger pixel area. In order to see how the k-ratio affects the retrieval process, equation (8) can be reformulated as

\[
\frac{[\rho^t(\mu_1, \mu, \phi, \lambda) - \rho^a_n(\mu_1, \mu, \phi, \lambda)]}{T^t(\mu_1, \mu, \phi, \lambda) - T^a_n(\mu_1, \mu, \phi, \lambda)} = k. \tag{29}
\]

It is evident that the value of k affects the results of a retrieval strongly. If the k-ratio is computed using values that are simply averaged, values that are not representative for any of the pixels in the superpixel area could be most certainly encountered. For example, consider an area where half of the larger area is covered with pixels having a high and the other half having a low k-ratio. When the k-ratios are averaged the end result would be wrong for the ADV method. Furthermore, as both of the AATSR views are employed, in simple averaging of reflectance one cannot be certain that mutual nadir/forward pixels are used when the k-ratio is determined. This could lead to situations where, in principle, nadir and forward view reflectance come from different pixels.

The chosen approach to average measured reflectance is to find those pixels that are most representative for an area, and at the same time are mutual to nadir and forward views. This is achieved by using the following method:

1. At least 50 % of the pixels belonging to the superpixel area must pass the cloud screening tests. This step ensures that enough information is present for the following steps.

2. Produce a histogram of reflectance measured in the 1.61 µm channel separately for nadir and forward reflectance. Typically seven bins are used ranging from zero to the maximum of the measured reflectance. The infrared channel is used here because the effect of aerosols is small and the effect of gases is negligible. That is, the measured reflectance is considered in first approximation to have only surface contribution.

3. Choose the nadir/forward bins that have the maximum number of reflectance values.

4. Find out which pixels that are in the chosen bins are mutual to nadir and forward views.

5. If there are more than ten values left, average the chosen reflectance values and use them in retrieval. If less than ten values are left, the surface reflectance in the area is considered to vary too much and retrieval is not executed.

The number of bins in the histogram determination is a compromise between loss of data and degeneration towards simple averaging. If too many bins were used, there would be too few pixels for the averaging of the reflectance. This situation would be potentially poor in statistical sense when only few pixels would represent the whole area. If too few bins were used, a too wide range of reflectance values would be accepted. This would allow pixels that could lead to a situation where the whole representative search of the k-ratio approach would become meaningless.
The over-ocean ASV algorithm utilizes the above described reflectance averaging for the sake of uniformity. A simple average could be used also for the ASV, as it can be assumed that the ocean surface reflectance is quite smooth over the 0.1° × 0.1° superpixel area.

The other test for the averaged reflectance measures the sufficiently uniform atmosphere condition. The standard deviation of reflectance at 0.555 μm is used as a measure for the uniformity. The 0.555 μm channel is utilized here as it is sensitive to both aerosol and cloud conditions. If the standard deviation is too large for a superpixel, results are judged to be unreliable. Retrieval is still done and the results include the standard deviation which can then be applied by the end user to exclude unreliable areas. This test can be seen as an additional spatial cloud screening but can also invalidate the superpixel in a case when large aerosol gradients occur, such as in the presence of strong sources.

4 Ash-specific retrieval

In this section we describe how the standard ADV/ASV algorithm (CCI version) needs to be adapted when used for ash specific retrievals. One of the key aspects is that instead of global retrievals of clear sky aerosols, the retrieval is made only over ash affected pixels. The ash mask is described in section 5. In addition, the approach used in global retrievals with four basic aerosol types is not applicable to volcanic ash plumes: aerosol models specific for ash need to be used. As a first approximation, a simplified approach with two ash aerosol models is used. The models have the same refractive index, but different size distribution. The retrieval algorithm then finds the best mixing ratio between these two components, and the total number of particles (reference AOD level).

The main difficulties when applying the general ADV/ASV algorithm to aerosol optical depth retrieval over ash plumes are listed below:

1. The aerosol models and LUTs need to be modified. The selection of appropriate LUTs appears to be the major challenge in this approach, since only two (or three) models can be used. A more laborious alternative would be to modify the algorithm so that it could select from a range of effective radii (instead of just mixing two aerosol types with predefined radii).

2. Standard cloud screening cannot be used. This leads to problems when we try to retrieve ash over a cloud layer. See section 5.

3. The k-ratio method may not be valid for thick ash plumes over land, as it assumes, in first approximation, that the presence of aerosol does not affect the reflectance at 1.6 µm.

4. The ash plume height is not accounted for in the retrieval. The aerosol models assume an aerosol layer height of 0-2 km. Also, the ASV algorithm uses the forward view, but the ash mask uses the nadir, which may cause collocation problems.

5. The AOD needs to be converted to column mass load values, and the aerosol mode fractions need to be converted to effective radius $r_{eff}$. The value of ash density used is a major source of error.

These points are discussed in more detail in the following.

4.1 Aerosol models

The ADV/ASV algorithm is designed for retrieving the total aerosol load due to various natural or anthropogenic sources. This is achieved by using specific aerosol models for four basic types, and retrieving the best fitting mixing ratios in addition to the reference AOD level (or total number concentration $N$). The size distribution and refractive index of each aerosol model are set a priori. In the volcanic ash perspective, however, the dominating aerosol type is known to be the volcanic ash, but the size distribution and optical properties are unknown and evolve during transport. In the typical volcanic ash retrieval algorithms, often operating in the thermal infrared (TIR) wavelengths, the approach is to retrieve the best fitting combination of reference AOD level (or $N$) and the aerosol effective radius $r_{eff}$.

Instead of adopting the TIR method where LUTs are calculated with the same $m$ for various effective radii $r_{eff}$, we modify the general ADV/ASV approach only slightly, as a first approach. For simplicity, we use only two pre-defined models with the same $m$ but different $r_g$, and let the algorithm optimize the mixing ratio between these two models. From the retrieved mixing ratio we can then derive the effective radius (see Eq. 47 below). We use this simplified model in version 1.0 of the algorithm; mixing with more effective radii, or with different refractive indexes, may be tested in the following versions. We could also test using the effective radius as a decision parameter (instead of mixing ratio), but this would require more profound changes to the algorithm.

Instead of one of the four aerosol components shown in Table 2, we use a combination of the components specified in Table 3 in v1.0 ash-specific retrieval. The refractive index corresponds to volcanic andesite (Pollack et al., 1973), and the size distribution is calculated as in Eq. 9. In version v1.0 retrievals the aerosol types TH0 ($r_g = 1.0$ µm) and TI20 ($r_g = 2.0$ µm) were used. The selection of the mode radii depends on the wavelengths used in the retrieval. The visible wavelengths are sensitive to smaller particles than in the TIR approach. An analysis of the sensitivities is needed.
Table 3: The aerosol models used in the first version of ash-specific retrievals. Note that the aerosol layer height (ALH) is not appropriate for volcanic ash. More suitable models will be used in the future versions.

<table>
<thead>
<tr>
<th>model</th>
<th>$r_g$ ($\mu$m)</th>
<th>$\sigma$</th>
<th>$m$</th>
<th>ALH (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>TI01</td>
<td>0.142</td>
<td>1.700</td>
<td>1.47 - 0.0015i</td>
<td>0-2</td>
</tr>
<tr>
<td>TI10</td>
<td>1.0</td>
<td>1.700</td>
<td>1.47 - 0.0015i</td>
<td>0-2</td>
</tr>
<tr>
<td>TI20</td>
<td>2.0</td>
<td>1.700</td>
<td>1.47 - 0.0015i</td>
<td>0-2</td>
</tr>
</tbody>
</table>

4.2 $k$-ratio for ash plumes

The $k$-ratio approach is not expected to be valid for thick ash plumes over land. The $k$-ratio method assumes that the 1.6 $\mu$m TOA reflectance is unaffected by the aerosols, in the first approximation. For thick ash plumes this is obviously not true, and part of the aerosol signal may than be misinterpreted and removed as surface reflectance contribution. This may lead to significant underestimation of AOD over land (see e.g. Figs 15 and 16). The ASV algorithm over ocean does not use the $k$-ratio method, and may provide more reliable results. In any case, only limited information can be retrieved for opaque plumes; with the current algorithm AOD values higher than approximately 4 are not retrieved.

4.3 Aerosol layer height

The aerosol layer height (ALH) affects the results in two ways. First, it affects the radiative transfer calculations via the aerosol model: an aerosol layer height needs to be specified for each aerosol model. The values currently used (for historic reasons), with aerosol layer between 0 and 2 km height, are not suitable for typical ash plumes. The sensitivity of the radiative transfer calculations to the ALH should be studied: if necessary, the plume top height estimate could be used to select the correct ALH to be used in each retrieval. Second, since the ADV/ASV algorithm uses information from both AATSR views, the plume height causes collocation issues. In the $k$-ratio method (ADV), the nadir and forward view TOA reflectances are compared for a ground-level-collocated pixel.

In the ASV algorithm used over ocean the forward view is employed, in order to avoid sun glints. However, the ash mask, which is the same over land and ocean, uses only the nadir view. The nadir view ash mask gives the correct geolocation for the plume, whereas a forward view mask would be shifted due to the parallax. This causes collocation problems, since the elevated ash plume does not coincide with the ash mask in the forward view. If a separate ash mask is used for the forward view, a different BTD threshold would be needed. Also, the location of the forward view ash mask and AOD results would require a geolocation correction. Alternatively, the nadir view could be used in the ASV retrieval, with the pixels affected by the sun glint removed (or flagged as unreliable).

4.4 Mass loading

In this section we describe how the retrieved products, the mixing ratio $b$ and the AOD $\tau$ are converted to effective radius $r_{eff}$ and the column mass load $m_l$, using the size distributions assumed in the aerosol models. The log-normal size distribution form is fixed by two parameters, the mode radius $r_g$ and the standard deviation $\sigma$ (see Eq. (9) and Table 2), while the scale (total number of particles) is fixed by the retrieved AOD (or reference AOD level). The total size distribution is obtained by mixing the two components (see Table 3), but we will first consider a single component case for simplicity.

AOD, $\tau$, is defined as

$$\tau = \int_0^L \sigma_e(z)\,dz = L\sigma_e,$$

(30)

where the last form follows if we assume that the extinction coefficient $\sigma_e$ does not change within a layer of thickness $L$, and is zero elsewhere. The extinction coefficient is

$$\sigma_e = \pi \int r^2 Q_e(\lambda, r)n(r)\,dr,$$

(31)

where $Q_e(\lambda, r)$ is extinction efficiency (obtained from a Mie code), $n(r)$ is the differential number concentration, and the total number concentration is $N_0 = \int n(r)dr$. Inserting this expression in (30) gives,

$$\tau = \pi L \int r^2 Q_e(\lambda, r)n(r)\,dr.$$

(32)

The aerosol concentration (total mass per volume) is

$$C = \frac{4\pi}{3}\rho \int_0^\infty r^3 n(r)\,dr,$$

(33)
where \( \rho \) is the ash density, taken to be 2600 kg/m\(^3\) (Neal et al., 1995). The aerosol mass load (column mass, \([m]\)=kg/m\(^2\)) is

\[
m_l = C_L = \frac{4\pi}{3} \rho L \int_0^\infty r^3 n(r) \, dr = \frac{4}{3} \rho r^{\lambda}(\frac{\lambda}{Q}) \int_0^\infty r^2 Q_e(\lambda, r) n(r) \, dr,
\]

(34)

where in the last equality we have inserted \( L \) from (32). If we define an ‘effective extinction efficiency’ \( Q_{\text{eff}} \) by

\[
Q_{\text{eff}} = \int_0^\infty r^2 Q_e(\lambda, r) n(r) \, dr
\]

(35)

and use the definition of effective radius

\[
r_{\text{eff}} = \frac{\int r^2 n(r) \, dr}{\int r^2 n(r) \, dr},
\]

(36)

we can rewrite Eq. (34) as

\[
m_l = \frac{4}{3} \rho r^{\lambda} \frac{r_{\text{eff}}}{Q_{\text{eff}}(\lambda)}.
\]

(37)

Note that \( Q_{\text{eff}} \) as defined in (35) is not the same as \( Q_e(r_{\text{eff}}) \).

### 4.4.1 Number size distribution

We see that the mass loading \( m_l \) depends linearly on the AOD \( \tau(\lambda) \) in Eq. (37). The conversion parameter \( r_{\text{eff}}/Q_{\text{eff}} \) depends on the form of the number size distribution, but not on the total number concentration (i.e. reference AOD level). To see this, let us look at the number size distribution in more detail.

We use the log-normal distribution,

\[
\frac{dN(r)}{d\ln r} = \frac{N_0}{\sqrt{2\pi} \ln \sigma} \exp\left(-\frac{(\ln r - \ln r_g)^2}{2 \ln^2 \sigma}\right),
\]

(38)

where \( N_0 \) is the aerosol (total) number concentration, \( r_g \) is the mode radius, and \( \sigma \) is the standard deviation. The relation to the differential number size distribution \( n(r) \) used in the formulas above is given by

\[
n(r) = \frac{dN}{dr} = \frac{1}{r} \frac{dN}{d\ln r} = \frac{N_0}{\sqrt{2\pi} \ln \sigma} \exp\left(-\frac{(\ln r - \ln r_g)^2}{2 \ln^2 \sigma}\right).
\]

(39)

The qualitative form of the number size distribution is fixed by \( r_g \) and \( \sigma \), but for quantitative use we need to know the total number concentration \( N_0 \). This is obtained from the retrieved AOD value \( \tau \). We can define a reduced distribution \( n'(r) = n(r)/N_0 \), which does not depend on the AOD level. Similarly, we can define other reduced parameters, like

\[
\tau' = \tau/N_0 = \pi L \int r^2 Q_e(\lambda, r) n'(r) \, dr.
\]

(40)

We note that the effective radius \( r_{\text{eff}} \) and effective extinction efficiency \( Q_{\text{eff}} \) do not depend on \( N_0 \):

\[
r_{\text{eff}} = \frac{\int r^2 n(r) \, dr}{\int r^2 n(r) \, dr} = \frac{\int r^2 n'(r) \, dr}{\int r^2 n'(r) \, dr}.
\]

(41)

Using a Mie code, we can calculate the factors \( r_{\text{eff}} \) and \( Q_{\text{eff}} \) for each aerosol type, and store them in a look up table (LUT). Then the conversion from AOD to mass load is simply a matter of multiplication with the given numbers.

### 4.4.2 Mixture of two aerosol types

The ADV algorithm uses (at least) two aerosol LUTs, typically one for coarse and one for fine mode particles. The retrieval chooses an optimal mixing ratio of the two types for each pixel online. In this case, we use the same aerosol (andesite) but with two different size distributions. We need to generalize the discussion above for this combined size distribution. In this simple approach it is easy to calculate the new distribution \( n(r) \) and all the related parameters, but now we need to use the correct mixing ratio, which varies with each pixel. Thus the mass loading retrieval must be reconstructed. Luckily all the related parameters are linear, and thus it suffices to calculate the factors \( r_{\text{eff}} \) and \( Q_{\text{eff}} \) separately for the two distributions, and combine them when the mixture ratio is known.

Here we assume that linear mixing can be used,

\[
n(r) = b_1 n_1(r) + b_2 n_2(r),
\]

(42)
where \( n_1(r) \) and \( n_2(r) \) are the size distributions for the two aerosol types, \( b_i \) is the fraction of the component \( i \) of the mixture and \( b_1 + b_2 = 1 \). Instead of mixing number concentrations, we could mix AODs, volumes or masses. Linear mixing for all quantities is not always sufficient. For example in ADV the reflectance due to aerosols in Eq. (3) is computed using the modified linear mixing method.

![Figure 5: Bimodal distribution](image)

Figure 5: Bimodal distribution \( r_g = 0.142 \) (µm), \( \sigma = 1.7 \) and \( r_g = 1.0 \) µm, \( \sigma = 1.7 \) with three different mixing ratios. We see that the effective radius \( r_{\text{eff}} \) does not decrease much when we increase the fraction of the smaller component.

Now we can calculate, for instance,

\[
 r_{\text{eff}} = \frac{\int r^3 n(r) dr}{\int r^2 n(r) dr} = \frac{\int r^3 [b_1 n_1(r) + b_2 n_2(r)] dr}{\int r^2 [b_1 n_1(r) + b_2 n_2(r)] dr}. \tag{43}
\]

For calculating effective radius \( r_{\text{eff}} \) and column mass \( m_l \) for a known aerosol mixture (aerosol components and mixing ratio \( b \)) and known AOD, we need three factors (integrals) for each aerosol type. These are

\[
 F_p^{(1)} = \int_0^\infty r^2 n_p'(r) \, dr, \tag{44}
\]
\[
 F_p^{(2)}(\lambda) = \int_0^\infty r^2 Q_\alpha(\lambda, r) n_p'(r) \, dr, \tag{45}
\]
\[
 F_p^{(3)} = \int_0^\infty r^3 n_p'(r) \, dr, \tag{46}
\]

where the subscript \( p \) refers to the aerosol component, and \( n'(r) = n(r)/N_0 \) means the reduced distribution. These values are calculated off-line and stored to LUTs (see Table 8). Now we can write the effective radius \( r_{\text{eff}} \) as

\[
 r_{\text{eff}} = \frac{b_1 F_1^{(3)} + b_2 F_2^{(3)}}{b_1 F_1^{(1)} + b_2 F_2^{(1)}}. \tag{47}
\]

and the effective extinction efficiency \( Q_{\text{eff}} \) as

\[
 Q_{\text{eff}}(\lambda) = \frac{b_1 F_1^{(2)} + b_2 F_2^{(2)}}{b_1 F_1^{(1)} + b_2 F_2^{(1)}}. \tag{48}
\]

The column mass load is

\[
 m_l = \frac{4}{3} \rho \tau(\lambda) \frac{b_1 F_1^{(3)} + b_2 F_2^{(3)}}{b_1 F_1^{(2)}(\lambda) + b_2 F_2^{(2)}(\lambda)}. \tag{49}
\]

Here we see that \( m_l \) depends on \( \lambda \) via three terms, \( \tau(\lambda), F_1^{(2)}(\lambda), \) and \( F_2^{(2)}(\lambda) \). Of course, the actual ash particle mass in the column does not depend on the wavelength, and the numerical calculations show only weak dependence of \( m_l \) on the wavelength.

### 4.5 Error characterization

Error estimation for the main ADV and ASV product, AOD, was given in sections 3.1.3 and 3.2.4. The AOD uncertainties for each wavelength are given in the algorithm output. For the derived products, mass load and effective radius, separate uncertainties are not calculated in version 1.0. The main uncertainty in these derived products is the choice of the two aerosol models used in the retrieval, i.e. the refractive index \( m \), geometric radius \( r_g \), and the ash density \( \rho \). Naturally, the mass load results will vary significantly depending on the aerosol models. Further work on the subject is needed.
5 Ash Detection

5.1 Brightness temperature difference threshold

A volcanic ash plume can be identified using the difference in brightness temperatures in infrared channels centered at 11 $\mu$m and 12 $\mu$m ($T_{11}$ and $T_{12}$). The difference $\text{BTD} = T_{11} - T_{12}$ is positive for water droplets, water vapor and ice particles, but negative for silicate (SiO$_2$) particles (Prata, 1989). An example is shown in Fig. 6 for the Eyjafjallajökull eruption in 2010. For clear sky BTD is small but positive (Saunders et al., 1988). Many large ash plumes can also be visually detected from false color RGB AATSR images using the visible channels, as seen in Fig. 1. The ash plumes have a distinct color and shape (at least near the volcano point source), so that they can often be distinguished from meteorological clouds. Detection based on the visible channels is also studied (see section 5.4).

A critical parameter in the ash detection is the threshold for BTD, which is not necessarily always exactly zero. The observed TOA brightness temperatures depend on the amount of ash in the atmospheric column and the surface temperature. A significant hindrance is caused by presence of water vapor or water clouds, since these tend to change BTD in the opposite direction (Prata and Grant, 2001; Yu et al., 2002). Mixed water/ash clouds severely complicate classification of ash plumes and clouds. Because of the surface temperature difference between land and ocean, separate thresholds may be needed for ocean and land pixels. Also, there is difference in nadir and forward view thresholds due to different light path length, but only the nadir view is typically used for detection.

![Figure 6: Ash detection with different satellite instruments and preliminary FLEXPART data (Stohl et al., 2011), May 15th 2010 around noon, Eyjafjallajökull eruption. The AATSR data are obtained with the threshold BTD < 0 K, while for the MODIS data we have used BTD < −0.2 K. The SEVIRI data are obtained using more advanced ash detection scheme (Prata, 2013). The SEVIRI and FLEXPART preliminary data are obtained from the VAST test database. The AATSR/MODIS swaths are indicated by the yellow/orange areas (time indicated by the blue text). The red lines indicate CALIPSO overpasses (time indicated by the red text).](image-url)

Detection based on a threshold BTD is a trade-off between avoiding false alarms and acquiring sufficient data for any conclusions. For some cases, e.g. studying the transport of ash, it might be useful to use less stringent limits to get more information of where ash might be encountered. In some cases, stricter limits are
needed in order to eliminate perpetual false alarms. In version 1.0 of the algorithm, the threshold \( \text{BTD} < 0 \, \text{K} \) is systematically used.

### 5.2 Cloud Screening

In the ash plume studies, retrievals will only be made for pixels identified as ash. These pixels may also be affected by meteorological (e.g. water or ice) clouds. The standard cloud tests presented in section 3.3 cannot be used as such, since they tend to flag the ash plumes as clouds. In particular, the elevated ash plumes are cold, so the gross cloud test 1 cannot be used in ash specific retrievals. The AOD results will be unreliable for pixels affected by water clouds, i.e. for cases where there is a water cloud under the ash plume.

The retrieval algorithm has an option to mask out ash pixels possible contaminated by water/ice clouds, by employing a reflectance cloud test at the 659 nm channel. The test is based on TOA reflectance, and on an assumption that the brightest pixels in a scene are due to water clouds, rather than ash plumes. The test analyzes an approximately \( 512 \times 480 \) pixel scene at a time, creates a reflectance histogram, and automatically assigns a reflectance threshold, above which pixels are flagged as clouded (González, 2003). The cloud test is not fully reliable, and its use in the retrieval is optional. An example of removing clouded pixels is shown in Fig. 7.

![Figure 7: Eyjafjallajökull eruption 15 th May 2010. The false color RGB image (left) shows white water/ice clouds mixed with the grayish ash plume. The AOD values for the ash flagged pixels are saturated at the maximum level 4 due to cloud contamination. The data screened by the 659 nm reflectance cloud test (right) has approximately 30% less data points than the original data (center).](image)

### 5.3 Shadow mask

At a given position along the satellite track, the forward view may be obstructed by high plumes earlier on the track (Fig. 8). The nadir and forward views will then see different parts of the ash plume, and the retrieval is likely to produce erroneous results. The AATSR correlation method (ACM) height estimate algorithm can be run in parallel with the AOD retrieval (Virtanen and de Leeuw, 2013). ACM can detect and flag pixels where the forward view is obstructed. This 'shadow mask' can be used in post-processing to remove possibly erroneous data.

![Figure 8: Principle of the plume shadowing. A high feature can block the forward view (for lower features) in the along track direction. A reliable AOD value cannot be retrieved for pixels in the shadowed area. To clarify the terminology we emphasize that the 'shadows' discussed here have nothing to do with sun light being blocked, and we do not consider the solar zenith angles here. By 'shadow' we merely refer to areas where a high plume feature blocks the forward view for a certain distance in the along track direction.](image)

### 5.4 Reflectance threshold method

In the false color RGB images the volcanic ash plumes stand out as brownish or yellowish 'clouds' and can be distinguished from (water) clouds, land, and ocean. (In the false color RGB images in Fig. 1 'red' is obtained from 865 nm channel, 'green' from 659 nm, and 'blue' from the 555 nm channel.) Therefore, it should be possible to implement an algorithm for detecting the ash plumes using the three or four shorter wavelength channels...
as well. Here we describe a simple reflectance threshold method, which is still in testing phase and requires manual tuning of the thresholds. These methods are not used in version 1.0 of the algorithm. An example of using reflectance thresholds is shown for the Etna eruption on October 28th, 2002 in Fig. 9 below.

Figure 9: Reflectance thresholds: a) first the reflectance ratio \( r_1 = \frac{R_{865}}{R_{659}} \) is limited between 1.0 and 1.15 (ocean is below 1, land above 1.15). b) Then \( R_{865} \) is limited to \( R_{865} < 0.15 \) (ash is dark). c) Finally, the ratio \( r_2 = \frac{R_{659}}{R_{555}} \) is limited to \( 1 < r_2 \) (get rid of some coastal and cloud pixels). The remaining points follow more or less the ash plume (as judged by the BTD limit).

Next, we consider the eruption of Chaiten in Chile in May 2008. A thick eastward ash plume can be seen over Argentina on May 5th. There are obvious differences between the Chaiten (rhyolitic) and Etna (basaltic) ash characteristics. The reflectance thresholds used for Etna do not work for the Chaiten case as such, limits have to be adjusted. A similar approach gives the five area types shown in Fig. 10 a). For area type flagging we use the following conditions: if \( T_{11} - T_{12} < -0.7 \), pixel is flagged as ash (red); if \( 1.15 < \frac{R_{865}}{R_{659}} \), the pixel is flagged as land (green); if \( 0.99 < \frac{R_{659}}{R_{865}} < 1.01 \) or \( 0.3 < R_{659} \), pixel is cloudy (orange); if \( 0.02 < \frac{R_{555}}{R_{659}} \), pixel is ocean (light blue); and if none of the above holds, pixel is flagged as unknown (dark blue). The conditions are applied in such a way that the first condition that a pixel meets determines the flag.

Figure 10: May 5th 2008, Chaiten eruption. Area types; ash (red), land (green), cloud (orange), ocean (light blue) and unknown (dark blue) (see text). The other two panels show reflectance dependence on wavelength for different types, and the same normalized to \( R_{555} \).

Although there is some success in detecting the ash using this method, the thresholds have been hand picked for a best fit with the 'known' ash distribution obtained from the condition \( \Delta T < 0 \). The thresholds are case dependent, and hard to obtain automatically. Also, these thresholds are always balancing between excluding ash points and allowing for false points, and there is no clear physical bases for these thresholds, unlike for the temperature difference threshold.

5.5 False Alarms

There are several situations when negative TOA BTD is observed in the absence of volcanic ash. In the historic test cases it is often easy to discriminate between 'true' and false 'alerts', but for near real time operational use these false alerts may be a nuisance. Since the AATSR is no longer operational, not much emphasis is placed on improving the ash detection and the simple BTD threshold is used for ash detection in the first version of the algorithm. However, for the possible future use of the algorithm with the Sea and Land Surface Temperature
Radiometer (SLSTR), scheduled to be launched on Sentinel-3 in 2014, methods for removing the false alarms based on the VIS/NIR channels are studied. The height estimate may also be used in refining the ash mask in future versions of the algorithm.

**Desert Dust**

Desert dust is also detected and misidentified as volcanic ash by the temperature difference condition $\text{BTD}<0$. The ash signals seen over Sahara and in the Middle East in Fig. 11a are probably false alarms due to desert dust. Similar false alarms are seen over and near the Atacama desert in the case of Chaiten eruption. Desert dust is often bright and yellowish, so it should be possible to discriminate between ash and dust using the visible and NIR channels. A method to discriminate between ash and dust needs to be developed.

**Small negative BTD over land and coastlines**

On several occasions the $\text{BTD}<0$ threshold detects ash near Arctic coastlines in April-May 2010, when no ash is present over open ocean or land. These apparent false alerts appear both over ocean and land (Fig. 11b) near the coastlines, particularly in night time retrievals. These anomalies are probably due to fog, arctic haze or other such phenomena. Negative BTD is also observed over elevated terrain, such as the Alps or the Himalayas (Fig. 11c), in absence of ash or dust.

![Figure 11: a) Day time ash detection in May 2010, color indicating the date. Anomalous 'ash plumes' observed near the northern coast lines, and desert dust misidentified as in Sahara and Middle-East. b) Negative BTD values observed over land in Scandinavia in the absence of any 'visible' plume. c) Negative BTD over Himalayas.](image)

**False-negatives**

It is possible that the ash detection algorithm incorrectly flags an ash-affected pixel as free of ash, i.e. produces a false-negative flag. With the BTD threshold method this may happen for instance when the ash plume is optically thick (opaque), and the TOA brightness temperature is nearly the same for both 11 and 12 $\mu$m channels. Another cause for false-negative ash flags can be high water-vapor load, which acts in the opposite direction, i.e. increases BTD. No water vapor corrections (WVC) are used in the present version of the algorithm.

**5.6 Night Time Detection**

We can use the thermal channels of AATSR to detect the ash plumes in night time orbits. Although the AOD retrieval using visible channels cannot be done, we still get information on the spatial and temporal distribution of the ash. The use of night time orbits doubles the overpass frequency over a specific scene, in principle.
A Appendix: List of AATSR test cases

In Table 4 below we list 33 volcanic ash test cases for which daytime AATSR overpasses were found, with ash flagged pixels. Note that this is not a complete list and more cases can be found e.g. for Eyjafjallajökull. Table 5 gives the corresponding AATSR filenames used in retrievals. We also show automatically generated figures of the AOD for the test cases. The results are not filtered for water clouds, thus high AODs are often seen.

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Table 4: AATSR VIS/NIR volcano cases.

As an example, we show some automatically generated statistics of the AOD retrieval for the Eyjafjallajökull eruption in Fig. 12 and in Table 6. Similar data is produced for all eruptions (not shown).

Figure 12: Time series of daily average AOD, column mass and effective radius, Eyjafjallajökull 2010. For the mass load and effective radius the thick blue line is calculated from all single pixel values, with the dotted blue lines indicating the uncertainty (±σ), while the dashed red line shows averages calculated after filtering with default thresholds (see Virtanen and de Leeuw (2013) for details).
Table 5: AATSR VIS/NIR volcano cases; AATSR orbit file names.

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<td>Etna 2002</td>
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</table>

Table 6: Daily average values (standard deviations) of AOD and related parameters, Eyjafjallajökull 2010.
Figure 13: AOD at 555 nm for Eyjafjallajökull. The yellow areas show AATSR swaths, with the blue text indicating the time. The average AOD (standard deviation) and number of ash flagged pixels for each day (N) are given in the legend.
Figure 14: AOD at 555 nm for Grímsvötn.

Figure 15: AOD at 555 nm for Puyehue-Cordón Caulle.
Figure 16: AOD at 555 nm for Chaiten.

Figure 17: AOD at 555 nm for Kasatochi and Merapi (right).

Figure 18: AOD at 555 nm for Etna.
B Appendix: List of Acronyms

AATSR  Advanced Along Track Scanning Radiometer
ADV    AATSR dual view (algorithm)
ALH    aerosol layer height
AOD    Aerosol optical depth
ASV    AATSR single view (algorithm)
ATBD   Algorithm theoretical basis document
BTD    Brightness temperature difference
FMI    Finnish Meteorological Institute
IR     Infra-red
LUT    Look-up-table
NIR    Near-infrared
NRT    Near-real time
SEVIRI Spinning Enhanced Visible and Infrared Imager (aboard MSG)
SLSTR  Sea and Land Surface Temperature Radiometer
SST    Sea surface temperature
SWIR   Short wave thermal infrared
TIR    Thermal-infrared
TOA    Top of atmosphere
UV     Ultra-violet
VAST   Volcanic Ash Strategic-initiative Team
VIS    Visible (wavelengths)
VZA    Viewing (satellite) zenith angle

C Appendix: Data product format

The FMI ADV/ASV/ACM VIS/NIR ash retrieval algorithm version 1.0 produces output in text format (ASCII), and contains a lot of data fields mainly for testing and developing purposes. The output contains data for both AOD retrieval (ADV/ASV) and for the height estimate (ACM). The final product(s) will be given in netcdf files, the content of which will be decided later. The output file is named as yyyymmddhhhh_0x.dat, where yyyy, mm, dd, and hhhh describe the year, month, day and starting time of the AATSR orbit. Each row of the text file corresponds to a single pixel, and each column corresponds to a different data field. The data consist of 52 data fields as described in Table 7.
<table>
<thead>
<tr>
<th>Column</th>
<th>Name</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>lon</td>
<td>Longitude</td>
</tr>
<tr>
<td>2</td>
<td>lat</td>
<td>Latitude</td>
</tr>
<tr>
<td>3</td>
<td>AOD555</td>
<td>Aerosol optical depth at 555 nm.</td>
</tr>
<tr>
<td>4</td>
<td>AOD659</td>
<td>Aerosol optical depth at 659 nm.</td>
</tr>
<tr>
<td>5</td>
<td>AOD865</td>
<td>Aerosol optical depth at 865 nm (only over ocean).</td>
</tr>
<tr>
<td>6</td>
<td>AOD1600</td>
<td>Aerosol optical depth at 1.6 µm.</td>
</tr>
<tr>
<td>7</td>
<td>uncert.555</td>
<td>AOD uncertainty at 555 nm.</td>
</tr>
<tr>
<td>8</td>
<td>uncert.659</td>
<td>AOD uncertainty at 659 nm.</td>
</tr>
<tr>
<td>9</td>
<td>uncert.865</td>
<td>AOD uncertainty at 865 nm.</td>
</tr>
<tr>
<td>10</td>
<td>uncert.1600</td>
<td>AOD uncertainty at 1.6 µm.</td>
</tr>
<tr>
<td>11</td>
<td>BTD</td>
<td>Brightness temperature difference (T_{11} - T_{12})</td>
</tr>
<tr>
<td>12</td>
<td>height</td>
<td>Main (single-pixel) height estimate from 11 µm thermal band (km asl).</td>
</tr>
<tr>
<td>13</td>
<td>C</td>
<td>Cross correlation coefficient, a measure of the height estimate quality.</td>
</tr>
<tr>
<td>14</td>
<td>std_c</td>
<td>Standard deviation of C with respect to all allowed pixel shifts.</td>
</tr>
<tr>
<td>15</td>
<td>along t. shift</td>
<td>Along track pixel shift of the forward view image.</td>
</tr>
<tr>
<td>16</td>
<td>across t. shift</td>
<td>Across track pixel shift needed for plume top collocation.</td>
</tr>
<tr>
<td>17</td>
<td>av. height</td>
<td>Moving average of single-pixel heights. Used to smoothen the noisy height data.</td>
</tr>
<tr>
<td>18</td>
<td>std_av</td>
<td>Standard deviation of the height in the moving averaging window (MAW), in km.</td>
</tr>
<tr>
<td>19</td>
<td>n_av</td>
<td>Number of pixels in the MAW. Only ash-flagged pixels are used in the average.</td>
</tr>
<tr>
<td>20</td>
<td>nad_mu</td>
<td>Satellite (viewing) zenith angle, nadir view.</td>
</tr>
<tr>
<td>21</td>
<td>nad_mu0</td>
<td>Sun zenith angle, nadir view.</td>
</tr>
<tr>
<td>22</td>
<td>nad_relazi</td>
<td>Relative azimuth angle between sun and satellite, nadir view.</td>
</tr>
<tr>
<td>23</td>
<td>for_mu</td>
<td>Satellite (viewing) zenith angle, forward view.</td>
</tr>
<tr>
<td>24</td>
<td>for_mu0</td>
<td>Sun zenith angle, forward view.</td>
</tr>
<tr>
<td>25</td>
<td>for_relazi</td>
<td>Relative azimuth angle between sun and satellite, forward view.</td>
</tr>
<tr>
<td>26</td>
<td>day</td>
<td>(Time stamp) the ENVISAT day, calculated from 1/1/2000.</td>
</tr>
<tr>
<td>27</td>
<td>second</td>
<td>(Time stamp) the ENVISAT second, calculated from midnight.</td>
</tr>
<tr>
<td>28</td>
<td>AOD_1 555</td>
<td>AOD due to the first aerosol component (fine mode) at 555 nm.</td>
</tr>
<tr>
<td>29</td>
<td>AOD_2 555</td>
<td>AOD due to the second aerosol component (coarse mode) at 555 nm.</td>
</tr>
<tr>
<td>30</td>
<td>R_{659} (TOA)</td>
<td>Top-of-atmosphere reflectance at 659 nm for cloud screening, nadir view.</td>
</tr>
<tr>
<td>31</td>
<td>land/sea flag</td>
<td>Land or sea flag: 0 for ocean, 1 for land.</td>
</tr>
<tr>
<td>32</td>
<td>k-ratio</td>
<td>Ratio between the forward and nadir view reflectance.</td>
</tr>
<tr>
<td>33</td>
<td>aerosol 1 ratio</td>
<td>Proportion of the first aerosol component in the mixture (values 0-1).</td>
</tr>
<tr>
<td>34</td>
<td>LM-info</td>
<td>Quality flag from the Levenberg-Marquardt optimization algorithm.</td>
</tr>
<tr>
<td>35</td>
<td>T_{11}^N</td>
<td>Nadir view brightness temperature at 11 µm.</td>
</tr>
<tr>
<td>36</td>
<td>T_{12}^N</td>
<td>Nadir view brightness temperature at 12 µm.</td>
</tr>
<tr>
<td>37</td>
<td>T_{11}^F</td>
<td>Forward view brightness temperature at 11 µm.</td>
</tr>
<tr>
<td>38</td>
<td>T_{12}^F</td>
<td>Forward view brightness temperature at 12 µm.</td>
</tr>
<tr>
<td>39</td>
<td>height(2)</td>
<td>Single-pixel height estimate with medium size correlation window (km).</td>
</tr>
<tr>
<td>40</td>
<td>height(1)</td>
<td>Single-pixel height estimate with small correlation window (km).</td>
</tr>
<tr>
<td>41</td>
<td>distance</td>
<td>Distance from the volcano in question (km).</td>
</tr>
<tr>
<td>42</td>
<td>r_eff</td>
<td>Aerosol effective radius (µm).</td>
</tr>
<tr>
<td>43</td>
<td>mass</td>
<td>Aerosol column mass (mass concentration) (g/m^2).</td>
</tr>
<tr>
<td>44</td>
<td>cloud test 3</td>
<td>659 nm reflectance cloud test flag, 0 for cloud free, 1 for clouded.</td>
</tr>
<tr>
<td>45</td>
<td>cloud test 4</td>
<td>865/659 reflectance ratio cloud test. Not used.</td>
</tr>
<tr>
<td>46</td>
<td>σ_{cws}</td>
<td>Standard deviation of along track shift wrt correlation window size (CWS).</td>
</tr>
<tr>
<td>47</td>
<td>k_p</td>
<td>Number of acceptable pixels in correlation matrix.</td>
</tr>
<tr>
<td>48</td>
<td>shadow flag</td>
<td>Shadow flag. 0 for clear view, 1 for obstructed forward view.</td>
</tr>
<tr>
<td>49</td>
<td>best av. height</td>
<td>Best average height (km asl)^1.</td>
</tr>
<tr>
<td>50</td>
<td>σ_m</td>
<td>Standard deviation of across track shift in MAW.</td>
</tr>
<tr>
<td>51</td>
<td>i_line</td>
<td>AATSR grid line number; along track index.</td>
</tr>
<tr>
<td>52</td>
<td>i_cols</td>
<td>AATSR grid column (pixel) number; across track index.</td>
</tr>
</tbody>
</table>

1) Moving average calculated from filtered single pixel heights.

Table 7: Combined output of AATSR VIS/NIR AOD retrieval algorithm (ADV/ASV) and height estimate (ACM). Intermediate data, version 1.0.
D Appendix: LUTs used

The size distribution is described by $r_g$ and $\sigma$, refractive index is $m$, and the conversion parameters for $r_{\text{eff}}$ and $m_l$ are given by the integrals

$$F^{(1)}(r) = \int_0^\infty r^2 n'(r) dr, \quad F^{(2)}(\lambda) = \int_0^\infty r^2 Q_e(\lambda, r)n'(r) dr, \quad F^{(3)} = \int_0^\infty r^3 n'(r) dr. \quad (50)$$

<table>
<thead>
<tr>
<th>model</th>
<th>$r_g$ ($\mu$m)</th>
<th>$\sigma$</th>
<th>$m$</th>
<th>alh (km)</th>
<th>$\lambda$ ($\mu$m)</th>
<th>$F^{(1)}$ ($\mu$m$^2$)</th>
<th>$F^{(2)}(\lambda)$ ($\mu$m$^2$)</th>
<th>$F^{(3)}$ ($\mu$m$^3$)</th>
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</thead>
<tbody>
<tr>
<td>TI01</td>
<td>0.142</td>
<td>1.700</td>
<td>1.47 - 0.0015i</td>
<td>0-2</td>
<td>0.55</td>
<td>0.00450</td>
<td>0.008584</td>
<td>0.009311</td>
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Table 8: The aerosol models used in the first version of ash-specific retrievals. The integral $F^{(2)}(\lambda)$ depends on wavelength; values are given for the VIS/NIR wavelengths 0.555, 0.659, 0.865, and 1.61 $\mu$m.
References


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Yu, T., Rose, W. I. and Prata, A. J., Atmospheric correction for satellite-based volcanic ash mapping and