Tracking and quantifying volcanic SO2 with IASI, the September 2007 eruption at Jebel at Tair

Lieven Clarisse, Pierre-François Coheur, A.J. Prata, Daniel Hurtmans, A. Razavi, Tanguy Phulpin, Juliette Hadji-Lazaro, Cathy Clerbaux

To cite this version:

Lieven Clarisse, Pierre-François Coheur, A.J. Prata, Daniel Hurtmans, A. Razavi, et al.. Tracking and quantifying volcanic SO2 with IASI, the September 2007 eruption at Jebel at Tair. Atmospheric Chemistry and Physics, European Geosciences Union, 2008, 8 (24), pp.7723-7734. <10.5194/acp-8-7723-2008>. <hal-00349230>

HAL Id: hal-00349230
https://hal.archives-ouvertes.fr/hal-00349230
Submitted on 13 Jan 2016

HAL is a multi-disciplinary open access archive for the deposit and dissemination of scientific research documents, whether they are published or not. The documents may come from teaching and research institutions in France or abroad, or from public or private research centers.

L’archive ouverte pluridisciplinaire HAL, est destinée au dépôt et à la diffusion de documents scientifiques de niveau recherche, publiés ou non, émanant des établissements d’enseignement et de recherche français ou étrangers, des laboratoires publics ou privés.
Tracking and quantifying volcanic SO₂ with IASI, the September 2007 eruption at Jebel at Tair

L. Clarisse¹, P. F. Coheur¹, A. J. Prata², D. Hurtmans¹, A. Razavi¹, T. Phulpin³, J. Hadji-Lazaro⁴, and C. Clerbaux⁴,¹

¹Spectroscopie de l’Atmosphère, Service de Chimie Quantique Photophysique, Univ. Libre de Bruxelles, Brussels, Belgium
²Norwegian Institute for Air Research, Kjeller, Norway
³Centre National d’Études Spatiales, 18 Avenue E. Belin, 31401 Toulouse, France
⁴UPMC Université Paris 06, CNRS UMR 7620, Service d’Aéronomie/IPSL, Paris, France

Received: 14 July 2008 – Published in Atmos. Chem. Phys. Discuss.: 9 September 2008
Revised: 27 November 2008 – Accepted: 28 November 2008 – Published: 22 December 2008

Abstract. In this paper we demonstrate the potential of the infrared Fourier transform spectrometer IASI in analysing volcanic eruptions, using the September 2007 eruption at Jebel at Tair as an illustrative example. Detailed radiative transfer calculations are presented, simulating IASI-like transmittance spectra for a variety of volcanic plumes. We analyse the sensitivity of IASI to SO₂ at different altitudes and demonstrate that IASI is in principle capable of sensing SO₂ down to the surface. Using the brightness temperature difference of well chosen SO₂ channels as a filter, we are able to track the plume of the Jebel at Tair eruption for 12 days, on a par with state of the art UV sounders. A method is presented for quickly estimating the altitude of a volcanic plume based on the relative intensities of the SO₂ absorption lines. Despite recent advances, it is still very challenging to retrieve vertical profiles of SO₂ from nadir viewing satellites. Currently the most accurate profiles in nadir are retrieved using backtracking of the plume with atmospheric transport models. Via full inverse retrievals using the optimal estimation method, we show the possibility of extracting medium coarse vertical profiles from IASI data. The retrieval allows us to present an evolution of the total mass of SO₂ in the plume for the Jebel at Tair eruption. An analytical relation is derived between brightness temperature differences and concentrations, which fits the experimental data very well. The spectral range of IASI also allows retrieval of volcanic aerosols. In the initial plume of the Jebel at Tair eruption, volcanic aerosols were found in the form of ice particles, for which we derived particle sizes.

1 Introduction

Yearly, an estimated 5–20% of the total global sulphur emission is due to volcanic sulphur dioxide (Halmer et al., 2002). The impact of SO₂ on the atmosphere is very much dependent on the height of the eruption. In the troposphere, sulphur dioxide oxidises to sulphate after which it is washed out quickly. The lifetime of SO₂ in the stratosphere is of the order of weeks, during which time it gradually forms sulphuric acid aerosols, which have a lifetime in the atmosphere of about three years. These aerosols have an important impact on the global climate as they increase the earth’s albedo through scattering of sunlight. Quantifying natural SO₂ is crucial in comparing natural versus anthropogenic emissions and their respective impact on climate (Andres and Kasgnoc, 1998). Large eruptions also provide an opportunity to validate climate models because of the relatively short perturbation on the climate system (Robock, 2000). Apart from these long term interests, monitoring SO₂ has an application as a surrogate for tracking volcanic ash clouds for aviation hazards (Stunder et al., 2007).

Due to their spatial and temporal distribution, volcanic eruptions are best monitored from space. Traditionally, ultraviolet sounding was mostly utilized for detecting volcanic SO₂, while thermal infrared sounding was traditionally used for the detection of volcanic aerosols. That said, both IR and UV sounders are capable of detecting SO₂, volcanic aerosols (Krotkov et al., 1999; Wen and Rose, 1994) and volcanic ice (Rose et al., 2003). Spaceborne ozone monitors measure radiation using backscattered ultraviolet sunlight (BUV) and are able to measure the SO₂ absorption feature at 230–330 nm. Examples include the Total Ozone Mapping Spectrometer (TOMS, e.g. Krueger et al. (1995)), the Global Ozone Monitoring Experiment (GOME, e.g. Eisinger et al., 1998; Loyola et al., 2008) and more recently the Scanning
In this section we outline some spectroscopic features of transmittance spectra relevant to the estimation of the altitude of a volcanic plume. We have simulated radiance spectra for a variety of plumes with different peak altitude and SO$_2$ concentration, in order to see their impact on the spectra. The calculations were carried out with the software Atmosphit, developed at the Université Libre de Bruxelles (http://home.scarlet.be/dhurma/atmosphit.html). We start by summarizing the main features of Atmosphit’s forward modelling. For the details we refer to Barret et al. (2005a,b), Coheur et al. (2005) and Rodgers (2000).
3.1 Forward model

Dividing the atmosphere in different layers \( k \), with assumed constant temperature \( T_k \), pressure \( p_k \) and volume-mixing ratios, the upward radiance at wavenumber \( \nu \) entering layer \( k+1 \) can be approximated by

\[
R_{k+1}(\nu) = R_k(\nu)\tau_k + B(\nu, T_k)(1 - \tau_k),
\]

where the sum runs over different species \( i \). Here the coefficients \( n_{i,k} \) denote average molecular number densities and \( C_{i,k} \) molecular partial columns. In the evaluation of the latter quantity, the effects of refraction are incorporated via ray tracing. The coefficients \( \Phi_i \) represent the discrete absorption line shape or continuous band shape specific to the species \( i \). These are calculated using a Voigt line shape and the line parameters in the HITRAN database (Rothman et al., 2005). The absorption continua of water vapour, carbon dioxide, oxygen and nitrogen are incorporated using the MT_CKD 1.03 model (Mlawer et al., 2003). Finally the Fourier transform infrared instrumental lineshape of IASI, including apodization and field of view, is taken into account.

3.2 Simulations and discussion

The atmospheric conditions were specified using a standard tropical model (15N latitude) defining average temperature, pressure and volume mixing ratios for 28 molecules in one kilometre thick layers from the ground up to sixty kilometres height. The surface temperature was fixed at 300 K. For each simulation a different amount of SO\(_2\) was inserted in a single 500 m thick layer located between 1 km and 30 km. Figure 3 shows the difference in brightness temperature (BT)
between the baseline (300 K) and the three strongest absorption lines of the ν1 and the ν3 band as a function of the altitude of the SO2 plume. The specific lines with the strongest absorption vary with altitude and concentration of the plume and give an accurate idea of the sensitivity of IASI. The four different curves represent different amounts of SO2 in Dobson Units (1 DU equals 2.69 × 10^{16} molecules per cm^2). Both bands have highest sensitivity between 16 and 18 km, which coincides with the cold point tropopause in the tropical model. The ν3 band is clearly more sensitive than the ν1, except below 4 km. In the region 1300–1400 cm\(^{-1}\) the water vapour in the atmosphere absorbs nearly all radiation coming from the ground. The atmosphere is only sufficiently transparent at a higher altitude, where the concentration of water vapour is lower (Prata and Bernardo, 2007). For a 100 DU SO2 plume at 1 km altitude, the highest brightness temperature differences are 1 K in the ν1 band and 0.7 K in the ν3 band. IASI has an instrumental noise lower than 0.12 K and should be able to see this. IASI is therefore theoretically capable of sensing volcanic SO2 in the planetary boundary layer, but much will depend on the precise atmospheric conditions such as the water vapour profile and the thermal contrast between the surface and the first atmospheric layer.

Figure 4 illustrates the impact of water vapour on the observed line intensities. Figure 4a shows the pseudo transmittance spectra of an over-simplified atmosphere containing nothing but 100 DU of SO2 (again at different altitudes). The only aspect really affecting the intensity of the lines is the thermal contrast between the source (surface) and the temperature of the SO2 layer. Figure 4b shows the same pseudo transmittance spectra, but this time in an atmosphere populated with water vapour (and the other molecules from the standard model). As mentioned above, water vapour absorbs almost all radiation at low altitude, making the observation almost insensitive to SO2 in the boundary layer. The spectra in Fig. 4 are pseudo transmittance spectra because they represent the ratio of the BT spectra for an atmosphere with and without SO2. In contrast, the effective transmittance is only very weakly dependent on the altitude. To scan daily global data for exceptional concentrations of SO2, we have designed a BT difference filter which calculates BTs at four different IASI channels. The channels at ν′\(_0\)=1407.25 and ν′\(_0\)=1408.75 are used to estimate the baseline; ν\(_3\)=1371.50 and ν\(_3\)=1371.75 to estimate absorption in the ν3 band. These
channels were chosen so as to minimise false alerts caused by water vapour interference, while maximising sensitivity. On 30 September, BT differences as large as 50 K (for the \nu_3 band) were measured in the plume from Jebel at Tair. In view of Fig. 3 this gives a lower bound of 80 DU on the concentration and puts the peak altitude immediately in the range 12–21 km. For this particular eruption this is already a reasonably good estimate. Absolute intensities for eruptions where the plume does not reach a high altitude will not give us any of this information, as the BT difference will be much smaller.

To quantify the plume’s altitude, instead of looking at the absolute intensities, we can also look at the relative intensities of the absorption lines. As an example, Fig. 5 shows the ratio \( R(1347.25, 1368.00) \) between the pseudo transmittance coefficients in the IASI channels at 1347.25 cm\(^{-1}\) and at 1368.00 cm\(^{-1}\) as function of altitude and concentration. For large concentrations of SO\(_2\) this ratio tells us whether the peak altitude is below or above 10 km. For the eruption at Jebel at Tair, typical pseudo transmittance spectra on 30 September have ratios between 1.4 and 1.5, putting the plume in the range 14–19 km with concentrations higher than 60 DU. Analogous results were obtained with other choices of channels for the ratio \( R \). The pseudo transmittance spectra were calculated from the radiance spectra (the target spectra) through division by a reference spectrum. The reference spectra were chosen to have the same baseline as the target spectra without being affected by SO\(_2\). They are typically found geographically close to the target (see also Prata and Bernardo, 2007). It is very important that the reference spectrum is chosen properly, as the ratio \( R \) can be quite sensitive to a bad choice. Looking at BT differences and intensity ratios is a quick way of estimating concentration and altitude of a volcanic plume. In the case of large volcanic eruptions, more accurate estimates can be obtained via inverse SO\(_2\) profile retrievals, which we will discuss in detail in the next section.

4 Full retrievals

4.1 Inverse model

As with the forward model, the inverse model was also carried out with the help of Atmosphit. Again, we will sketch the method and refer for details again to Barret et al. (2005a), Barret et al. (2005b), Coheur et al. (2005) and Rodgers (2000). Let us assume for the time being that the forward model is linear, i.e.

\[ y = Kx + \epsilon, \]

with \( y \) representing the measurement vector (the radiances) and \( x \) representing the state vector (parameters we would like to retrieve such as surface temperature, and concentrations and constant model parameters such as the pressure and temperature profiles). The forward model is represented by the weighting function or kernel \( K \), the measurement error by \( \epsilon \).

This type of inverse problem is generally ill-posed, with no exact solution. There are a number of ways to find a regularized solution \( \hat{x} \) which approximates the true state \( x \). The optimal estimation method presented in Rodgers (2000) uses a Bayesian approach. It combines the measurement result \( y \) and its measurement covariance matrix \( S_y \) with the a priori knowledge of the state \( x_a \) and its measured covariance \( S_a \) to yield a maximum a posteriori solution \( \hat{x} \) with covariance \( \hat{S} \). The solution is given by (here we give the so-called \( n \)-form of the solution)

\[ \hat{x} = x_a + (K^T S_y^{-1} K + S_a^{-1})^{-1} K^T S_y^{-1} (y - Kx_a) \] (3)

and

\[ \hat{S}^{-1} = K^T S_y^{-1} K + S_a^{-1}. \]

Writing

\[ K = \frac{\partial y}{\partial x} \quad \text{and} \quad G = \frac{\partial \hat{x}}{\partial y}, \]
with the gain matrix \( G = (K^T S_a^{-1} K + S_a^{-1})^{-1} K^T S_a^{-1} \), we can introduce the averaging kernel

\[
A = G K = \frac{\partial \hat{x}}{\partial x}.
\]

In words, the weighting function is the sensitivity of the measurement to the true state, the gain matrix the sensitivity of the retrieved state to the measurement and the averaging kernel the sensitivity of the retrieved state to the true state. In the idealized case where the problem is well determined we have \( \hat{x} = x + K^{-1} \varepsilon \) and \( A = I \). More generally though, we can rewrite Eq. (3) as

\[
\hat{x} = A x + (I - A)x_0 + G \epsilon,
\]

so that the retrieved state is the weighted average of the true state with the a priori state. The more \( A \) resembles the identity matrix \( I \), the better \( \hat{x} \) will resemble \( x \). The resemblance is commonly quantified by the trace of \( A \) and goes under the name “degrees of freedom for signal” (DFS). It gives an estimate on the number of retrieved independent pieces of information.

For a moderately non-linear problem like ours, we have \( y = F(x) + \epsilon \). Via Newton methods, the solution can be approximated by the iterative solution

\[
\hat{x}_{i+1} = x_0 + (K_i^T S_i^{-1} K_i + S_i^{-1})^{-1} K_i^T S_i^{-1} [y - F(\hat{x}_i) + K_i(\hat{x}_i - x_0)],
\]

where \( K_i = \frac{\partial F}{\partial x} \) and \( K_i = K(x_i) \).

### 4.2 Retrieval parameters

For the Jebel at Tair eruption we did not use the \( \nu_1 \) band for the retrieval. The presence of an ice signature makes an accurate retrieval hard in this band (see Sect. 5). The band is also affected by emissivity features over arid land. The plume passed over both the Arabian desert and the Gobi desert, where the emissivity is strongly dependent on the wavenumber in the region 800–1300 cm\(^{-1}\) (see e.g. Li et al., 2007). For the retrievals we considered therefore only the \( \nu_3 \) absorption feature. The disadvantage of this band is the strong competition of water vapour absorption. An accurate retrieval of the vertical profile of water vapour is essential, and for this reason the fitting range was set to 1310–1450 cm\(^{-1}\). This range not only contains the complete \( \nu_3 \) band of SO\(_2\), but also contains a large part of the \( \nu_2 \) band of H\(_2\)O. Other interfering molecules in this range include CH\(_4\) and N\(_2\)O and total columns were fitted for those. For water vapour, concentrations were retrieved in 1 km thick partial columns, from the surface up to 20 km. For the first 36 h after the eruption, SO\(_2\) was retrieved in three partial columns: 12–15 km, 15–18 km and 18–21 km. This choice was motivated by the analysis presented in Sect. 3. For spectra recorded after 1 October, one total column was fitted between 15–18 km (see Sect. 4.3.2). Apart from the partial and total columns of the different species, the surface temperature has also been fitted. The spectra for the retrieval were selected on the basis of the BT difference between the \( \nu_3 \) band and the baseline. Only spectra with a BT difference larger than 0.5 K were retained. A priori values for all molecules were taken from either the tropical or midlatitude summer model, and pressure and temperature profiles were interpolated from data from the ECMWF to match date and location of the IASI measurements. The measurement covariance matrix was taken to be diagonal \( S_\epsilon = \sigma_\epsilon^2 I \), with \( \sigma_\epsilon = 2.5 \times 10^{-8} \text{W/(cm}^2\cdot \text{sr} \cdot \text{m}^{-1}) \), which is of the order of the RMS of a typical spectral fit (see Fig. 6). The a priori covariance matrix \( S_\eta \) was taken as Gaussian with an off-diagonal coupling constant of 1 km (see Barret et al., 2005a for the explicit formula). For SO\(_2\) the diagonal values of the covariance matrix were chosen proportional to the BT difference in the \( \nu_3 \) band. For the other molecules they were chosen to

---

**Fig. 6.** A sample fit of a spectrum from the first day of the eruption. The dotted lines correspond to \( \sigma_\epsilon = \pm 2.5 \times 10^{-8} \text{ W/(cm}^2\cdot \text{sr} \cdot \text{m}^{-1}) \).

**Fig. 7.** The retrieved vertical profile of SO\(_2\) corresponding to the spectrum displayed in Fig. 6.
Fig. 8. Different views of the volcanic plume from the eruption at Jebel at Tair on 30 September 2007 around 18:48 UTC. (a) Brightness temperature differences in the ν3 band, (b) Retrieved SO2 concentrations in DU, (c) Altitude of peak SO2 concentration, (d) Degrees of freedom for signal.

be constant and corresponding to a relative standard deviation of 25%. The iteration from Eq. (4) was repeated until $|F(\hat{x}_i) - F(\hat{x}_{i+1})| < 0.2\sigma_e$.

Note that background values of SO2 together with a covariance matrix with a very large diagonal and small off-diagonal elements comes close to using a more standard least-squares minimization. An a-priori and covariance matrix could be constructed using the vertical profiles of many different SO2 plumes. Such information is not available and would remain inadequate viewing the dramatic and unpredictable character of volcanic eruptions. The more general optimal estimation fit was used in this study, as this is the preferred method for the other interfering molecules and gives detailed information on the results (e.g. vertical sensitivity). The choice of taking the a priori variability proportional to the BT difference will be justified in the following section.

4.3 Retrieval of the SO2 profile

4.3.1 Day of the eruption

A typical fit and corresponding vertical profile for spectra on 30 September are shown in Figs. 6 and 7. Figure 8 summarises the total retrieval, i.e. the brightness temperature difference, the concentration, the DFS of the SO2 retrieval and the retrieved height of the concentration peak. The first thing to note is the clear correlation between the BT difference and the SO2 concentration (see in this context also Prata et al., 2003). Figure 9 shows the relation more explicitly. For low concentrations (<30 DU) there is an almost linear correlation between the two, while for higher concentrations the BT difference progressively saturates. This correlation can be derived using simplified versions of Eqs. (1) and (2). Let us assume a layer of SO2 centred at an altitude of 16.5 km. The radiance at wavenumber $\nu$ entering the layer can be written as $B(\nu, T_a)$, where $T_a$ is total brightness temperature of the layers below. The radiance leaving the layer can then be written as

$$B(\nu, T_b) = B(\nu, T_a)\tau + B(\nu, T_{165})(1 - \tau),$$

with $T_{165}=192$ K the temperature at 16.5 km according to the tropical model, $\tau = \exp(-c_1C)$ the transmittance, $C$ the SO2 concentration and $c_1$ a coefficient dependent on temperature, pressure and thickness of the layer. The BT difference we wish to find is given by $\Delta BT = T_a - T_b$ (in view of the
Correlation between the brightness temperature difference and the retrieved SO$_2$ concentrations. Each circle represents data from one spectrum, the solid line is given by $\Delta T = T_a - T_b$, with $T_b$ given in Eq. (5) with parameters $T_a=243$ K and $c_1=0.034$ DU$^{-1}$.

altitude, we ignore absorption above the SO$_2$ layer). After straightforward calculus one finds

$T_b = \frac{A}{\ln[1 + GH/(H\tau + G(1-\tau))]},$ (5)

with $A = hcv/k$, $G = \exp(A/T_a) - 1$, and $H = \exp(A/T_{165}) - 1$. From Eq. (5) one can see that when the concentration is very large the observed temperature $T_b$ approaches $T_{165}$. This again shows the importance of thermal contrasts between $T_a$ and $T_b$ as $\Delta BT$ is effectively bounded by $T_a-T_b$. Using linear squares regression, we found values of $T_a=243$ K and $c_1=0.034$ DU$^{-1}$. The resulting curve (Fig. 9) matches the data very well. Note that the a priori variability (covariance matrix) for the SO$_2$ retrievals was chosen proportional to the BT difference. The non-linear correlation between the concentration and the BT difference indicates that the retrievals were not affected by this choice. We verified on a sample of spectra that as long as it was chosen large enough, the retrieved concentration is pretty much independent of the variability.

The DFS graph in Fig. 9 shows a moderate correlation with the SO$_2$ concentration. The maximum DFS is around 2.3; and with a retrieval between 12 and 21 km this corresponds to a vertical resolution smaller than 4 km. It also shows that three partial columns is the right choice for the retrievals for the first day of the eruption. Fewer columns would mean disregarding available information, while more partial columns lead to erratic profiles because of the necessarily large diagonal of the a priori covariance matrix $S_a$. Another good indication is that for almost all retrieved pixels, the peak of the averaging kernel is the same as the corresponding retrieval level (i.e. the information on a partial column comes primarily from the right altitude). The altitudes where the SO$_2$ concentration reaches its peak in the profile is typically between 14 km and 17 km. The highest concentrations all had a peak altitude of about 16.5 km. This result is in excellent agreement with information from other satellites and atmospheric transport model simulations (Eckhardt et al., 2008).

4.3.2 Tracking the plume

As shown in Fig. 2, IASI was able to follow the plume for almost twelve days. Three partial columns of SO$_2$ were fitted from the spectra of the first 36 h after the eruption. Retrievals from spectra after that typically had a DFS of 1 and lower. For this reason, only total columns of SO$_2$ were retrieved for those. The evolution of the total mass is shown in Fig. 10. It was calculated using the average total column in an area around the observed plume. The first three days after the eruption, the total column peaks around 46 kt, after which it drops to around and below 10 kt. OMI estimated total SO$_2$ mass (assuming center of mass altitude 17 km and 5 km thickness) 57 kt, 43 kt, 31 kt and 24 kt on October 1 through 4, respectively, which fits well with exponential decay rate 3.4 days (see http://so2.umbc.edu/omi/pix/special/2007/redsea/altair07.php). Figure 10 shows that IASI estimates agree well with OMI tonnages on all days, except on 1 October, when the OMI mass was estimated 25% higher (the OMI overpass was before the beginning of the eruption on 30 September). The reason for disagreement on 1 October needs further investigation.

5 Ice

Ice can be a significant component of volcanic eruption clouds (Rose et al., 2003, 2004) and can hinder retrieval of volcanic ash as well as play a role in scavenging SO$_2$ (Texctor et al., 2003). Ice nucleation may also be a common
occurrence in volcanic clouds due to the seeding potential of volcanic ash particles (Durant et al., 2008). Measurements from the SEVIRI instrument reported by Eckhardt et al. (2008) suggest that ice was present in the Jebel at Tair eruption cloud. This was deduced by differencing top-of-atmosphere brightness temperatures in SEVIRI channels centred near 11 and 12 µm, which are strongly positive for semi-transparent ice clouds (Wu, 1987). The cause of the differences has been determined through radiative transfer theory (Parol et al., 1991) to be due to differential absorption and scattering by ice (spheres and cylinders have been considered), resulting ultimately from differences in the complex refractive indices of ice at the two wavelengths.

If ice particles were present in the Jebel eruption cloud, then in principle they should be detectable in the IASI brightness temperature spectra, especially within the window region between 800–1000 cm\(^{-1}\). To investigate whether ice was present, we have performed some discrete ordinates radiative transfer calculations for ice spheres using Mie theory, and coupled this with Modtran-4 radiance calculations for atmospheric absorption using a nearby radiosonde profile. The resulting radiances were converted to brightness temperatures and compared directly with an IASI spectrum in the eruption cloud. The real and imaginary parts of the refractive index of ice based on the values provided in Warren (1984) for the region 720–1450 cm\(^{-1}\), tabulated in wavelength (µm) at resolutions of 0.1 to 0.3 µm (10–30 cm\(^{-1}\) at 10 µm) were interpolated to the resolution of IASI of 0.25 cm\(^{-1}\). There is a peak in the imaginary part of the refractive index near 800 cm\(^{-1}\) and a decrease either side of this maximum. Thus absorption of infrared radiation would be expected to decrease as the wavenumber increases from 800 cm\(^{-1}\) to 1000 cm\(^{-1}\). The rate of this decrease in absorption depends on ice particle size and optical depth (Huang et al., 2004). To get an idea of the location of the possible ice plume, we have plotted in Fig. 11 the brightness temperature difference of two almost clear channels at 996.25 cm\(^{-1}\) and 923.75 cm\(^{-1}\) on 30 September. The location of the ice plume does not exactly coincide with the location of the SO\(_2\) plume, indicating a difference in altitude.

The IASI brightness temperature spectrum of a pixel in the cloud is shown in Fig. 12, together with the radiative transfer simulation. The simulation uses the interpolated refractive indices in a Mie calculation to determine the extinction coefficients due to scattering and absorption, the phase functions and asymmetry parameters for a range of particle sizes...
6 Conclusions

The spectral range of IASI covers both the ν1 and ν3 band of SO2 with a good signal to noise ratio. We have analysed IASI’s sensitivity to SO2, and found very good sensitivity for SO2 at high altitudes and under favourable conditions some sensitivity for SO2 in the boundary layer. Using the moderate eruption at Jebel at Tair as a test-case, we have demonstrated that the brightness temperature difference of some well chosen channels in the ν3 band can be used to track SO2 over a large period and large distance from its source. Depending on concentration, thermal contrast and altitude, also vertical profile information of a volcanic SO2 plume can be obtained. We have shown here that ratios of relative intensities of absorption lines in the ν3 band can be used to give estimates of the plume’s altitude. Both the tracking of the plume using the brightness temperature filter and the method of estimating the plume’s altitude are computationally straightforward. Since the IASI data is available in NRT these methods can be used for operational analysis. We have presented detailed SO2 profile retrievals using the optimal estimation method, giving us estimates of the concentration and peak altitude. We have shown that the total SO2 burden agrees well with OMI measurements. IASI also allows direct detection of volcanic ash and ice and the high spectral resolution allows the retrieval of microphysical properties such as particle sizes. A detailed study of aerosol detection with IASI will be the subject of further research.

Acknowledgements. IASI has been developed and built under the responsibility of the Centre National d’Études Spatiales (CNES, France). It is flown onboard the Metop satellites as part of the EUMETSAT Polar System. The IASI L1 data are received through the EUMETCast near real time data distribution service. L. Clarisse and P. F. Coheur are respectively Scientific Research Worker (Collaborateur Scientifique) and Research Associate (Chercheur Qualifié) with F.R.S.-FNRS. C. Clerbaux and J. Hadji-Lazaro are grateful to CNES for scientific collaboration and financial support. The research in Belgium was funded by the F.R.S.-FNRS (M.LS. n°4511.08), the Belgian State Federal Office for Scientific, Technical and Cultural Affairs and the European Space Agency (ESA-Prodex arrangements C90-219). Financial support by the “Actions de Recherche Concertées” (Communauté Française de Belgique) is also acknowledged. The authors would like to thank Jean-Noël Thépaut for providing us with the data of the ECMWF wind fields which were used to produce Fig. 2. The authors are also grateful the anonymous referees and N. Krotkov for their valuable suggestions and corrections.

Edited by: A. Richter

References


Rose, W. I., Bluth, G., and Watson, I.: Ice in volcanic clouds: When and where?, in: Proc. of the 2nd Int. Conf. on Volcanic Ash and Aviation Safety, OFCM, Washington, D.C., Session 3, 61, Ex-
tended observations of volcanic SO$_2$ and sulfate aerosol in the stratosphere, 2004.


