MERIS Surface Pressure

Algorithm Theoretical Basis Document
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1 Introduction

The retrieval of surface pressure from satellite based observations is of great interest due to its potential to fill the gaps of surface based measurement networks. Representing one of the main driving quantities of the weather on earth, a precise knowledge of the global distribution of surface pressure is an important input parameter for weather models. The remote sensing of surface pressure from space is on the other hand a highly ambitious task because of the precision needed: In order to be of any use as model input data, the accuracy of a retrieval algorithm should be in the range of 1 hPa (≈ 0.1% at 1000hPa, Bengtsson (1979)). The air pressure at a given location on the earth surface is determined by the mass of air in the vertical column above. In order to measure this mass from space, the “differential absorption technique” can be applied, which is based on exploiting spectral gaseous absorption features. The absorbing gas has to be well mixed in the atmosphere and moderately absorbing in the spectral range under consideration in order to allow the determination of the air mass traversed by the detected photons. Using a band with too much absorption results in saturated absorption bands that do not carry any information about photon path length, whereas a weak absorption band does not provide enough sensitivity. Therefore, the oxygen A-band at 762nm is perfectly suited for a retrieval of surface pressure, since

1. oxygen is moderately absorbing at this particular spectral range, is well mixed and has a constant concentration in the atmosphere.
2. the absorption band is undisturbed by absorption of other atmospheric species or emission of the atmosphere or the earth.

Various authors have outlined the potential usefulness of measurements in the oxygen A band for the retrieval of surface pressure (Barton and Scott (1986); Mitchell and O’Brien (1987); O’Brien and Mitchell (1989)). Breon and Bouffies (1996) have applied such a method to measurements of POLDER by using the deviation of the “apparent pressure”, derived on a clear sky assumption, as an indicator for cloud appearance. They found an accuracy of apparent pressure of 100hPa. O’Brien et al. (1998) found more promising results using an airborne grating spectrograph with high spectral resolution, directed towards sun glint areas above ocean. Using two channels with moderate spectral resolution (Δλ = 0.5nm) they found an accuracy of 20hPa, whereas a use of highly resolved spectra (Δλ = 0.05nm) allowed the determination of surface pressure with a precision of 1hPa. The difference in accuracy appears due to the fact that spectrally highly resolved measurements enable a separation of the fractions of signal that are caused by reflection at the surface and in the atmosphere, whereas a single channel in the oxygen A band only provides one piece of information, namely the average photon path length.

MERIS (Medium Resolution Imaging Spectrometer) onboard ENVISAT, primarily designed for ocean colour remote sensing, provides measurements in 15 channels between 0.4μm and 1.0μm (see table 1), three of them located in the vicinity of the O2 A band (MERIS channels 10, 11 and 12). Since there is only one channel within the absorption band and MERIS observations do not provide sufficient information about aerosols and the temperature profile, the achievable
accuracy of a retrieval of surface pressure is limited. Therefore, this work does not aim at deriving surface pressure with sufficient accuracy to serve weather models, but demonstrating the performance of a state-of-the-art retrieval algorithm based on existing satellite data. Future satellite missions like EnMap (Stuffer et al. (2007)), OCO-2 (Crisp et al. (2004)), CarbonSat or FLEX will provide spectrally higher resolved measurements in the oxygen A-band and thus allow more accurate retrievals of surface pressure.

2 Background

MERIS is a programmable, medium-spectral resolution, imaging spectrometer (Rast et al. (1999)). It is one of ten core instruments on the polar orbiter ENVISAT (Environmental Satellite, launched on March 1st, 2002) flying at 800km in a sun-synchronous orbit with an equator crossing time of 10:30AM, descending node, and 98.5° inclination. MERIS consists of 5 identical pushbroom imaging spectrometers operating in the solar spectral range (390nm to 1040nm), arranged in a fan shape configuration, which covers a total field of view of 68.5° and spans a swath width of around 1150km. The spectral dispersion is achieved by mapping the entrance slit of a grating spectrometer onto a CCD array. The integration time, instrument optics and CCD array resolution are adjusted such that MERIS has a spatial resolution of 260m by 300m and a spectral sampling of 1.25nm. The instrument electronic data rate provides 15 channels, which are programmable by ground command in width and in position. In the regular operation mode the spatial resolution is reduced by a factor of 4 along and across track (reduced resolution mode). In the full resolution mode, the full spatial resolution is transmitted. The central wavelengths of the spectral channels as listed in Table 1 vary slightly across the field of view of MERIS. This “spectral smile” is caused by curvature of the image of the slit formed in the focal plane array, resulting in viewing angle-dependent central wavelengths of the spectral MERIS channels. In order to accurately determine the spectral smile of MERIS, spectral calibration campaigns are conducted repeatedly, using the full possible spectral resolution in the oxygen A band and solar Fraunhofer lines (Delwart et al., 2007).

One of the major limitations of the MERIS instrument is missing observations in the SWIR spectral domain. This poverty hamper pixel identification, especially in the presence of clouds. The difference between the derived surface pressure and the ECMWF surface pressure is identified as a good indicator of cloudy or non-cloudy pixels.

In the MERIS pixel identification process such a procedure is applied. For further details see (Brockmann et al., 2011).

<table>
<thead>
<tr>
<th>Band</th>
<th>Center (nm)</th>
<th>Width (nm)</th>
<th>Main usage</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>412.5</td>
<td>10</td>
<td>Yellow substance, turbidity</td>
</tr>
<tr>
<td>2</td>
<td>442.5</td>
<td>10</td>
<td>Chlorophyll</td>
</tr>
<tr>
<td>3</td>
<td>490</td>
<td>10</td>
<td>Chlorophyll, pigment</td>
</tr>
<tr>
<td>4</td>
<td>510</td>
<td>10</td>
<td>Suspended matter, turbidity</td>
</tr>
<tr>
<td>5</td>
<td>560</td>
<td>10</td>
<td>Chlorophyll, suspended matter</td>
</tr>
<tr>
<td>6</td>
<td>620</td>
<td>10</td>
<td>Suspended matter</td>
</tr>
</tbody>
</table>
Algorithm Description

The algorithm for the retrieval of surface pressure described herein is based on the exploitation of the absorption of solar radiation by oxygen at 0.76 μm. The strength of absorption can be related to the traversed absorber mass, since the transmission decreases as the traversed absorber mass increases. The so-called “differential absorption technique” is implemented in a wide field of remote sensing algorithms for the estimation of masses, e.g. the retrieval of atmospheric water vapour or trace gases. Since oxygen is well mixed in the atmosphere, the measured mass of oxygen is strongly related to the average photon path length in the atmosphere. In case of clear sky measurements above bright land surfaces, the vast majority of photons detected by MERIS stems from reflection at the surface, enabling the retrieval of surface pressure.

The quantity carrying the information about the photon path length is the atmospheric transmission at 0.76 μm, which can not be measured directly but is approximated by the ratio $r$ of an absorbing channel within the oxygen $A$ band (MERIS band 11 at 761.875nm) and an artificial non-absorbing measurement at the same spectral location:

$$ r = \frac{L_{11}}{L_{\text{window}}} $$

The virtual non-absorbing measurement $L_{\text{window}}$ is approximated by interpolating two window channels below and above the oxygen $A$ band (MERIS bands 10 at 753.75nm and 12 at 778nm, see table 1), accounting for the spectral slope of the surface albedo.

The surface pressure retrieval algorithm is based on radiative transfer simulations using the Matrix Operator Model (MOMO; Fischer and Grassl (1984); Fell and Fischer (2001)). The simulations, covering all combinations of influencing parameters within the physically reasonable bounds, were used to derive coefficients of a multi-dimensional non-linear regression, relating the measured radiance to the surface pressure. The regression approach was chosen in order to obtain an algorithm, which is able to operate in near real time without needing a high amount of calculation power and main memory. Artificial neural networks have proven to be suitable tools to perform the regression.

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<table>
<thead>
<tr>
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<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>7</td>
<td>665</td>
<td>10</td>
<td>Chlorophyll</td>
</tr>
<tr>
<td>8</td>
<td>681.25</td>
<td>7.5</td>
<td>Chlorophyll</td>
</tr>
<tr>
<td>9</td>
<td>708.25</td>
<td>10</td>
<td>Atmospheric correction, ’red edge’</td>
</tr>
<tr>
<td>10</td>
<td>753.75</td>
<td>7.5</td>
<td>Cloud optical thickness, cloud-top pressure reference</td>
</tr>
<tr>
<td>11</td>
<td>761.875</td>
<td>3.75</td>
<td>Cloud-top / Surface pressure</td>
</tr>
<tr>
<td>12</td>
<td>778</td>
<td>10</td>
<td>Aerosol, vegetation</td>
</tr>
<tr>
<td>13</td>
<td>865</td>
<td>20</td>
<td>Aerosol, atmospheric correction</td>
</tr>
<tr>
<td>14</td>
<td>885</td>
<td>10</td>
<td>Water vapour reference</td>
</tr>
<tr>
<td>15</td>
<td>900</td>
<td>10</td>
<td>Water vapour</td>
</tr>
</tbody>
</table>

Table 1: Central wavelength and bandwidth (fwhm) of MERIS
3.1 MOMO

Assuming a plane-parallel atmosphere, media of any optical thickness with any vertical inhomogeneity as well as any spectral resolution can be considered by MOMO, whereas three-dimensional effects cannot be simulated. An advanced k-distribution technique is used to incorporate gaseous absorption (Goody and Young (1989); Bennartz and Fischer (2000)). These calculations were based on the HITRAN 2004 data set (Rothman et al. (2005)), providing shape and strength parameters of the single absorption lines of the main atmospheric gases.

3.2 Algorithm specification

3.2.1 Artificial neural network

Artificial neural networks (ANN) are used in a wide field of remote sensing applications. If trained and applied properly, ANN algorithms are fast and accurate tools for the inversion of satellite measurements. The artificial neural network used for inverting MERIS radiances to surface pressure values is a multi layer perceptron (MLP) with three layers and 150 hidden neurons. Using MOMO simulations as training data bases, three ANN-versions were developed for different temperature regimes, namely tropical, subarctic summer and US standard profiles. These model atmospheres (McClatchey et al. (1972)) have been widely used in the atmospheric research community and provide standard vertical profiles of pressure, temperature, water vapour and ozone density. The version implemented in the MERIS ground segment is based on a US standard atmosphere.

3.2.2 ANN input specification

The input to the artificial neural network is composed of
1. the radiance in MERIS band 10,
2. the radiance ratio \( r \) of MERIS band 11 (stray light corrected) and the window radiance interpolated from bands 10 and 12,
3. the aerosol optical thickness at 550nm (fixed to 0.15!)
4. the cosine of the solar incident angle
5. the cosine of the viewing angle
6. the cosine of the azimuth distance (viewing azimuth – solar azimuth, \( 0^\circ = \text{sensor opposite of sun} \)) times the sinus of the viewing angle.
7. the central wavelength of MERIS band 11.

The output of the ANN is the surface pressure above land.

The definition range of the input parameters is defined in table 2
### Empirical stray light correction and central wavelength of band 11

In the operational processing of L1-data, MERIS radiances are corrected for stray light within the instrument. However, significant differences in the pressure retrievals between the individual cameras of MERIS and in-camera gradients are found, hinting at a not properly working correction scheme. Therefore an empirical stray light correction was developed by FUB. The correction scheme is based on the optimization of a simple stray light model by fitting the retrieval of surface and cloud-top pressure to reference data for several selected scenes. The reference data was constructed from digital elevation models combined with ECMWF sea level pressure for clear sky scenes and MSG-brightness temperatures for cloudy scenes. A detailed description of the optimization process and the results can be found in Lindstrot et al. (2010).

### Sensitivity Studies

In order to determine the sensitivity of MERIS measurements to the individual influencing quantities, radiative transfer calculations were performed, using again the radiative transfer model MOMO. The simulations were analyzed with respect to the influence of surface pressure and other geophysical parameters like aerosol optical thickness and scale height, surface albedo and the temperature profile. Additionally, the impacts of instrumental parameters (spectral channel position and width) were determined.
4.1 Sensitivity to surface pressure, aerosol properties and instrumental parameters

Since the sensitivities to the examined parameters depend on the surface brightness and the aerosol loading, four different cases were studied:

a) an optically thin aerosol layer above a dark land surface (AOT = 0.1, \(\alpha=0.2\)),

b) an optically thick aerosol layer above a dark land surface (AOT = 0.8, \(\alpha=0.2\)),

c) an optically thin aerosol layer above a bright land surface (AOT = 0.1, \(\alpha=0.6\)),

d) an optically thick aerosol layer above a bright land surface (AOT = 0.8, \(\alpha=0.6\)).

The sensitivities were determined for a tropical atmosphere, a surface pressure of 1000hPa and an aerosol scale height of 2km, assuming an exponentially decreasing profile of extinction and a non-absorbing aerosol. For all four cases, the relative sensitivity of \(r\) to the examined parameters was determined with respect to an increase of surface pressure of 10hPa (\(\xi_{SP}\)), an increase of AOT of 0.2 (\(\xi_{AOT}\)), an increase of aerosol scale height of 1km (\(\xi_{ASH}\)), a shift of spectral channel position of 0.1Å (\(\xi_{cw}\)) and an increase of channel width of 0.1Å (\(\xi_{cwp}\)). These values were chosen, because they cause a change of signal in the same order of magnitude.

Figure 1 shows the resulting sensitivities as percentage of \(R\) for all four cases. The sensitivity to surface pressure \(\xi_{SP}\) is shown by the stars, whereas the squares and crosses represent the spectral channel sensitivities \(\xi_{cw}\) and \(\xi_{cwp}\), respectively (shown exclusively for case a), since there is no dependence on the cases). The triangles and diamonds correspond to the aerosol sensitivities \(\xi_{ASH}\) and \(\xi_{AOT}\), respectively. Several conclusions can be drawn from figure 1:

- The sensitivity of \(r\) to a change of surface pressure has its maximum (\(\approx 1\%/10hPa\)) at about 762nm, which is close to the nominal wavelength of MERIS band 11. There is only a weak dependence of \(\xi_{SP}\) on the aerosol loading and the surface albedo.

- \(r\) is very sensitive to a change of channel position \(cp\) and channel width \(cw\), regardless of the atmospheric state. Below 761nm, a shift of 0.1nm of \(cp\) towards longer wavelengths is equivalent to a change of surface pressure of about 50hPa. An increase of \(cw\) here corresponds to a decrease of surface pressure of 20 - 40 hPa. Both \(\xi_{cp}\) and \(\xi_{cw}\) vanish in the region of 762nm, due to the balancing effect of the weakly absorbing Q-branch of the oxygen absorption band. Towards longer wavelengths \(\xi_{cp}\) has a positive sign, whereas \(\xi_{cw}\) is close to zero. The strong sensitivities and their variability across the absorption band emphasize the need of an accurate spectral calibration. The nominal center wavelength of MERIS band #11 was initially set to 761.875nm to avoid artefacts caused by inaccurate spectral calibration. However, the actual center wavelength varies roughly between 761 and 763nm, due to the spectral smile effect.

- The sensitivity to aerosols strongly depends on the case under consideration. Above bright surfaces (cases c and d)), both \(\xi_{AOT}\) and \(\xi_{ASH}\) are negative, indicating that an increase of aerosol optical thickness or scale height causes an increase of absorption. In case of \(\xi_{AOT}\), this is caused by longer photon paths due to the "trapping" of photons between surface and the aerosol layer. The small negative effect of an increase of scale height (leaving the optical thickness unchanged) is caused by the increased distance covered by the photons between the scattering events. Above dark surfaces (cases a and b)), an increase of scale height by 1km has a strong effect for a high optical thickness (case b)), due to the shielding of
absorption in the lower atmosphere. At 762nm, it corresponds to a decrease of surface pressure by 15hPa. The effect is on the order of -2hPa for a low optical thickness (case a)).

Figure 1: Sensitivity of channel ratio \( r \) to surface pressure (\( \Delta S_P = 10hPa \)), aerosol optical thickness (\( \Delta AOT = 0.2 \)), aerosol scale height (\( \Delta ASH = 1km \)), channel position (\( \Delta c_p = 0.1\AA \)) and channel width (\( \Delta c_w = 0.1\AA \)), depending on channel position of MERIS band 11. Sensitivities shown for nadir view and solar zenith of 36°.

The influence of the aerosol optical thickness seems to be hard to predict, since it changes both sign and magnitude for each case. This is due to a dependence of \( \xi_{AOT} \) on the surface albedo, as shown in figure 2.
Above dark surfaces like ocean ($\alpha \leq 0.05$), an increase of AOT at a scale height of 2km or 4km causes longer photon paths and a decrease of $r$, since the scattering at aerosols occurs below the Rayleigh scattering ($h_{\text{Rayleigh}} = 8\,\text{km}$). For a moderate surface albedo ($\alpha \leq 0.3$) an increase of AOT causes shorter photon paths, because of the shielding of absorption by the aerosols. For bright surfaces ($\alpha > 0.3$), the mentioned mechanism of trapping of photons between surface and aerosols dominates, resulting in path extensions and a decrease of $r$. There are additional dependencies on the solar zenith angle and the aerosol scale height and optical thickness. Regarding the surface pressure retrieval, the uncertainty introduced by assuming average values for the aerosol optical thickness does not exceed 10hPa for a surface albedo $\alpha \geq 0.2$.

4.2 Sensitivity to the temperature profile

The oxygen $A$ band is composed of several hundred individual absorption lines, that are subject to pressure-and temperature-dependent broadening processes in the atmosphere. In the lower atmosphere, the dominant process is pressure broadening. The resulting line can approximately be described by the Lorentz line shape:

$$f (v - v_0) = \frac{\alpha_L}{\pi / ((v - v_0)^2 + \alpha_L^2)}$$

where $\alpha_L$ is the Lorentz half-width at half maximum, which is roughly proportional to the number of collisions per unit time (Petty (2006)):

$$\alpha_L \propto pT^{-1/2}$$

Obviously, an increase of temperature at constant pressure results in narrower, more intense
absorption lines, whereas a decrease of temperature causes broader lines with weaker centres. It is therefore important to examine the effect of different temperature profiles on the measured broadband transmission within the oxygen $A$ band. A simulation study was conducted using seven different standard temperature profiles as defined by McClatchey et al. (1972), namely a tropical, a mid-latitude summer and winter, a subarctic summer and winter and a US standard atmosphere (see Figure 3, left panel). In addition, a calculation using a US standard temperature increased by 1K at all height levels was performed. The middle and right panels of figure 3 show the resulting deviations of the channel ratio $r$ and the corresponding change of surface pressure for five profiles, relative to the US Standard atmosphere.

![Figure 3: Left panel: Assumed standard temperature profiles (US Standard, US Standard+1K, subarctic winter, subarctic summer, mid-latitude winter, midlatitude summer, tropical profile). Middle panel: Relative difference of channel ratio $r$ for used profiles, relative to US standard case, depending on height. Right panel: Equivalent change of surface pressure, relative to US standard case, depending on height. Sensitivities shown for nadir view and solar zenith of 40°, central wavelength $\lambda_{11} = 761.875\,nm$.]

At a height level of 1000hPa, a maximum deviation of 40hPa can be found for a subarctic winter profile. However, since in the subarctic region MERIS can hardly measure during wintertime due to the absence of sunlight, it is more reasonable to confine the analysis to the remaining profiles. As the deviation rises with pressure level due to the increase of traversed air mass, maximum errors of $\sim$20hPa can be found at sea level. In case the temperature profile is known within 1K, the errors do not exceed 5hPa, which is to be taken as an upper boundary as the deviation from the real profile will generally not be a constant offset at all height levels. In case the error in the temperature profile changes its sign at some height level, the resulting error in surface pressure cancels out at least in parts.

4.3 Conclusion

Measurements within the oxygen $A$ band show a high sensitivity to surface pressure. At a MERIS-like spectral resolution, the maximum sensitivity is found around 762nm. In the vicinity of this spectral location the errors caused by the uncertainty of the exact channel position and
width are small, the nominal center wavelength of MERIS channel 11 is therefore well defined. The uncertainty caused by the unknown aerosol loading and vertical profile is ≤ 10 hPa, except for cases of very high aerosol loading above dark surfaces. The temperature profile has to be known with high precision in order to prevent a large bias in retrieved surface pressure. A deviation of 1 K at all height levels results in errors ≤ 5 hPa. Assuming a fixed US standard temperature profile results in biases of up to ±20 hPa in extreme cases like tropical or arctic conditions. A further source of error is the unknown water vapour column amount, as the measurements are only sensitive to the dry air pressure. In the tropics, the partial pressure of water vapour can reach values of up to 8-10 hPa. Assuming an appropriate average humidity, the resulting errors in surface pressure are ≤4 hPa in extreme cases and lower otherwise. Both the uncertainty caused by the aerosol optical thickness and the water vapour amount can be limited by using the values retrieved from MERIS as implemented in the operational MERIS ground segment.

5 Algorithm performance

Two implementations of SP_FUB, assuming a tropical and subarctic summer atmosphere, respectively, were validated using digital elevation model data combined with ECMWF sea level pressure. For three case studies above Northern Africa, Southeastern Asia and Greenland, SP_FUB was found to retrieve the surface pressure without bias in case the average temperature profile of the scene matches well with the one used as input to the radiative transfer simulations serving as training data to the neural network. The root mean square deviation from the reference data was found to be within 8 -14 hPa, depending on the scene under consideration. This noise is mainly caused by undetected clouds, the variability of the temperature profile, aerosols and instrument noise.

In order to demonstrate the validity of the US Standard version of SP_FUB, an additional validation study was conducted using clear sky scenes above Asia, South Africa and North America. The results are compiled in figures 4-7, each showing the DEM pressure in the upper left, SP_FUB pressure in the upper right, a difference plot and a section plot (along the lines indicated in the upper figures) in the middle panels and histograms of the comparison in the bottom panels.

For the scenes investigated the bias of SP_FUB varies between 2hPa and -18 hPa, depending on the actual atmospheric profile. The bias-corrected root mean square error varies between 9 hPa and 14 hPa. This noise is caused partly by undetected clouds (see e.g. Northern region, figure 4) and cloud edges (see e.g. Western region, figure 6), respectively. The residual noise is again expected to be caused by temperature variations, aerosols and instrument noise.

The stray light correction seems to work reasonably well, although there are residual camera effects. However, these effects are small as compared to uncorrected retrievals.
Figure 4: Comparison of SP_{DEM} and SP_{FUB} for scene above the Tarim basin (China) on 5th of June 2004.
Figure 5: Comparison of SP\textsubscript{DEM} and SP\textsubscript{FUB} for a MERIS scene above Georgia / Alabama on 5\textsuperscript{th} July 2004.
Figure 6: Comparison of $SP_{DEM}$ and $SP_{FUB}$ for a MERIS scene above China (Yellow River valley) on 1st of September 2004.
6 Conclusions

The algorithm for the retrieval of surface pressure \(\text{SP}_{\text{FUB}}\) shows good results above bright land surfaces. The derived surface pressure values are bias-free if the assumed temperature profile is close to the mean profile of the scene, as the comparison with surface pressure maps constructed from ECMWF / SRTM data and ECMWF / GLAS data revealed. The \(\text{SP}_{\text{FUB}}\)-version to be implemented in MEGS assumes a US standard temperature profile. It was validated using the same approach as above, showing a similar accuracy. It is expected to show a positive bias in
colder regions and a negative bias in warmer regions. This behaviour was verified by the validation exercise. In extreme cases (tropics, arctic regions), this bias can exceed ±20hPa.

The retrieval noise, as found in the investigated scenes (8hPa ≤ rmsd ≤ 15hPa, depending on scene under consideration) can be well explained with undetected clouds, the uncertainty introduced by the unknown aerosol loading and profile and the exact temperature profile, as shown in section 4.

References


Brockmann, C., A. Ruescas, K. Stelzer, 2011: MERIS Pixel Identification - ATBD 2-17, ESA-ESRIN.


Product Name: Cloud-top pressure
Product Code: MERIS.RRGS
Product Level: Level 2
Description of Product: Surface pressure

Product Parameters:
Coverage: global
Packaging: [hPa]
Range: 1050 hPa to 300 hPa (for polar stratospheric clouds: 50 hPa)
Sampling: pixel by pixel

Resolution:
radiometric: Wm-2sr-1µm-1
spatial: 1.2km (0.3km)

Accuracy:
radiometric: 2-4% (within precision of calibration)
geophysical product: 20 hPa

Geo-location requirements: 1-4 pixels, depending on use of cloud-top pressure

Appended Data:
Earth location, Quality mask (i.e. residual of inversion process)

Frequency: 1 product per orbit

Size of Product: TBD

Additional Information:
Identification of bands used in algorithm: l=753.75 nm, l =760 nm
Assumptions on MERIS input data: None
Identification of ancillary and auxiliary data: surface albedo, stray light correction coefficients, MERIS band 11 central wavelength
Assumptions on ancillary and auxiliary data: TBD
Input from other ENVISAT instruments: None