**Pre-operational Sentinel-3 snow and ice products**

Algorithm Theoretical Basis Document

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**1.Introduction**

This document is aimed at the description of the theoretical basis of the algorithms to determine properties of snow and ice from Sentinel-3 observations.

The following topics are covered:

* Atmospheric correction
* Snow extent determination/snow mask
* Bare ice extent determination
* Dirty ice extent determination
* Snow and ice albedo retrieval (spectral and broadband)
* Bottom-of-atmosphere snow reflectance retrieval
* Pollution load retrieval
* Size of snow grains determination
* Snow specific surface area determination

# Overview

Snow is composed of ice crystals in contact with each other and surrounded by air. Snow can include 'impurities' such as dust, soot, algae (e.g., Skiles et al. 2018). Here, we refer to impurities as ‘pollution’. Snow can also contain liquid water. The volume concentration of snow grains is usually around 1/3 with 2/3 of the snow volume occupied by air (Proksch et al. 2016). The concentration of pollutants is often low, that is, below 100 ng/g especially in polar regions (Doherty et al. 2010).

The algorithms described here are dedicated to the retrieval of snow optical properties such as snow spectral and broadband albedo and also snow microstructure (snow specific surface area and effective optical grain size). We propose a snow mask based on the Normalized Difference Snow Index (NDSI) and a technique to retrieve the concentration of pollutants in snow, which is possible only for the cases with relatively heavy (above 1ppmv) pollution load (Warren 2013).

**2.1 Ocean and Land and Colour Instrument**

Ocean and Land and Colour Instrument (OLCI) is a 21 band spectrometer that measures solar radiation reflected by the Earth’s atmosphere and surface with a ground spatial resolution of 300 m (see Table 1). The OLCI swath width is 1270 km. OLCI is installed on both Sentinel-3A and Sentinel-3B satellite platforms operated by the ESA in service to the EU Copernicus programme. The Sentinel-3A,B orbit at 802 km altitude, 98.6 orbital inclination and a 10:00 UTC sun-synchronous equatorial crossing time.

## Generated Products

We arrived at the list of planned products (Table 2) as being both based on a theory that is considered mature and comprising products of need by the global snow modelling and Earth Observation community.

In addition to the products listed in Table 2, we also provide cloud mask derived using an approach not discussed in this ATBD. Most of retrievals are based on the measurements at 865 and 1020nm*,* where the influence of atmospheric light scattering and absorption processes on top-of-atmosphere signal as detected on a satellite over polar regions is weak.

|  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |
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| ***Table 1. Band characteristics of the SENTINEL-3 Ocean and Land Colour Instrument (OLCI)[[1]](#footnote-2)***   |  |  |  |  | | --- | --- | --- | --- | | ***Band*** | ***λ centre (nm)*** | ***Width (nm)*** | ***Function*** | | *Oa1* | *400* | *15* | *Aerosol correction, improved water constituent retrieval* | | *Oa2* | *412.5* | *10* | *Yellow substance and detrital pigments (turbidity)* | | *Oa3* | *442.5* | *10* | *Chlorophyll absorption max., biogeochemistry, vegetation* | | *Oa4* | *490* | *10* | *High Chlorophyll, other pigments* | | *Oa5* | *510* | *10* | *Chl, sediment, turbidity, red tide* | | *Oa6* | *560* | *10* | *Chlorophyll reference (Chlorophyll minimum)* | | *Oa7* | *620* | *10* | *Sediment loading* | | *Oa8* | *665* | *10* | *Chlorophyll (2nd Chlorophyll abs. max.), sediment, yellow substance/vegetation* | | *Oa9* | *673.75* | *7.5* | *For improved fluorescence retrieval and to better account for smile together with the bands 665 and 680 nm* | | *Oa10* | *681.25* | *7.5* | *Chlorophyll fluorescence peak, red edge* | | *Oa11* | *708.75* | *10* | *Chlorophyll fluorescence baseline, red edge transition* | | *Oa12* | *753.75* | *7.5* | *O2 absorption/clouds, vegetation* | | *Oa13* | *761.25* | *2.5* | *O2 absorption band/aerosol corr.* | | *Oa14* | *764.375* | *3.75* | *Atmospheric correction* | | *Oa15* | *767.5* | *2.5* | *O2A used for cloud top pressure, fluorescence over land* | | *Oa16* | *778.75* | *15* | *Atmos. corr./aerosol corr.* | | *Oa17* | *865* | *20* | *Atmos. corr./aerosol corr., clouds, pixel co-registration* | | *Oa18* | *885* | *10* | *Water vapour absorption reference band. Vegetation monitoring* | | *Oa19* | *900* | *10* | *Water vapour absorption/vegetation monitoring (max. reflectance)* | | *Oa20* | *940* | *20* | *Water vapour absorption, atmos./aerosol corr.* | | *Oa21* | *1 020* | *40* | *Atmos./aerosol corr.* | | |
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***Table 2. SICE: Snow and ice products***

|  |  |  |  |  |  |  |
| --- | --- | --- | --- | --- | --- | --- |
|  | **Snow product name** | **Short name** | **Units** | **Expected range** | **Maximum acceptable uncertainty in modeling** | **Optimum uncertainty** |
| 1 | Snow mask (based on NDSI) | SM | - | 0 - no snow, 1 -snow | <10% | 5 % |
| 2 | Spectral snow albedo (planar and spherical) | SA | - | 0.5-1.0 | <10% | 5 % |
| 3 | Broadband snow albedo  (planar and spherical) | BBA | - | 0.5-0.9 | <15% in the blue  source: GCOS (WMO, 2011) | <5% for all wavelength  source: GCOS (WMO, 2011) |
| 4 | Snow Specific  Surface Area | SSA | m2 kg-1 | 20-200 | <15% | 5 % |
| 5 | Snow grain size (diameter) | SGS | mm | 0.02-0.2mm | <15% | 5 % |
| 6 | Concentra-tion of pollutants | CP | ppmv | 0.1-10.0 | - | - |
| 7 | Normalized difference snow index | NDSI | - | - | - | - |
| 8 | Normalized difference bare ice index | NDBI | - | - | - | - |
| 9 | Pollution mask | PM | - | - | - | - |
| 10 | Bottom-of-atmosphere reflectance | BOAR | - | 0-1 | <15% | <4% |

*Table 3 notes*

1. The snow mask is based on NDSI. We do not provide the fractional snow cover. However, we provide the NDSI in the output of the algorithm, which can be used to estimate the snow cover.
2. The retrieval of spectral snow albedo is based on processing of OLCI data.
3. SSA is separate from grain size because it has different accuracy associated with differing field measurement approach.
4. BBA (planar and spherical) is provided for three spectral broad bands.
5. To determine BOAR, the top-of-atmosphere reflectance is corrected for atmospheric aerosol/molecular scattering and absorption effects.
6. In addition to the products listed in in Table 2, the following products are given in output:
7. We also derive the type of pollutants and provide flags for bare clean and polluted ice as discussed below.

## 2. 4 Assumptions

A number of assumptions are made that produce an algorithm set that have been shown to be robust. The simplified snow layer model represents snow as a:

1. Horizontally homogeneous plane parallel turbid medium;

2. Vertically homogeneous layer;

3. Semi-infinite layer. Therefore, there is no need to account for the reflective properties of underlying surface.

4. Close packed effects are ignored (although ice fraction is roughly 30%).

5. Geometrical optics can be used to derive local optical snow characteristics.

7. Impurities (dust, soot, etc.) are located external to ice grains.

8. The single light scattering angular pattern is spectrally neutral in the spectral range given in Table 1.

9. Only pixels completely covered by snow are considered, i.e., pixels with ice and/or partially snow pixels are ignored.

10. The effects of slopes and snow roughness are not accounted for.

The output is provided if the OLCI reflectance at 1020nm is larger than 0.1 and the derived diameter of grains is larger than 0.1mm. These numbers are given in the configuration file and can be changed, if needed.

**3. Snow and ice property retrievals**

**3.1 The mechanics of retrievals: a bird view**

Retrievals are approached in two ways, depending on a threshold in band 1.

*Clean snow retrieval approach*

First of all we check the reflectance in OLCI band 1. If it is larger than the *dynamic* threshold value (THV), it is assumed that the ground scene is covered by unpolluted snow (the majority of pixels in the terrestrial cryosphere). The THV is derived from the synthetic radiative transfer calculations for the assumed aerosol optical thickness at 550nm, which is given in the input of sice.f. Then we provide an index for the clean snow in the output and derive snow spectral albedo in the spectral range 0.3-2.4 micrometers using the two-parameter analytical equation as described by Kokhanovsky et al. (2019) under assumption that atmospheric effects can be ignored at OLCI channels 865 and 1020nm (not in other channels). The unknown two parameters are derived from OLCI reflectances at 865 and 1020nm. This simple approach to atmospheric correction has appeared to produce highly accurate snow spectral albedo in the range 0.4-1.02micrometers with deviations from ground measurements below 0.01-0.02. The same is true for the broadband albedo.

*Polluted snow retrieval approach*

The atmospheric correction for the polluted snow case (low value of OLCI reflectance at 400nm) is treated in two ways depending on the OLCI reflectance at 1020nm.

*Case 1*

If OLCI reflectance at channel 21 is above 0.5, the retrievals for polluted snow are based

1. on the OLCI measurements at bands 16 and 21 (respectively 865 nm and 1020nm wavelengths) and extrapolation to the longer wavelengths using the analytical equation for the spectral surface albedo identical to that as used for a clean snow.
2. on the OLCI measurements at the wavelengths below 865nm corrected for gaseous absorption and light scattering by aerosol in the framework of the theory described below (see Eq. (10)). The albedo inside O2 absorption band is derived using the linear interpolation of results for neighboring channels.

In this dark surface regime, we assume that scattering and absorption of light by surface impurities and atmosphere can be ignored at the wavelengths 865 and 1020nm. Such an assumption is similar to the approach used for the clean snow.

*Case 2*

In the case of dark snow and ice pixels and band 21 under 0.5 TOA reflectance, atmospheric correction of measurements *at all OLCI channels* must be performed. We can no longer assume that pollutants and other effects have small influence on OLCI reflectance above 865nm. Here, we use the transcendent Eq. (10) to account for gaseous absorption and light scattering by aerosol for all OLCI channels. We have found that such an approach produces good results outside oxygen and water vapor absorption bands. Therefore, the albedo inside O2 and water vapor absorption bands is derived using the linear interpolation of results for neighboring channels.

**3.2 Details of retrievals for the clean snow case**

The following relationship (Kokhanovsky et al., 2018) is used to retrieve the snow spherical albedo (see Appendix 1 for the definition) from the OLCI reflectance:

(1)

where

(2)

and we use the following approximation for the angular functions (Kokhanovsky et al., 2019)

(3)

Here, is the reflectance of underlying surface in assumption that there are no absorption processes in the medium under study (), s the cosine of the solar zenith angle, is the cosine of the viewing zenith angle. The relationship between the reflectance and spherical albedo given by Eq.(1) can be derived from the asymptotic radiative transfer theory (Kokhanovsky et al., 2019). This relationship allows for the determination of spherical albedo using reflectance observations at a given observation geometry. The procedure has been verified using airborne measurements of albedo and reflectance over a bright cloud field with the spherical albedo in the range 0.8-0.95 (Kokhanovsky et al., 2007). The plane albedo (see Appendix 1) can be easily derived from spherical albedo. Namely, it follows (Kokhanovsky et al., 2019):

(4)

The spherical albedo for clean snow can be presented in the following form (Kokhanovsky et al, 2018):

(5)

where *l* is the *effective absorption length (EAL)* in snow and α is the bulk absorption coefficient of ice*.* TheEAL *does not depend* on the wavelength in the OLCI spectral range as demonstrated by Kokhanovsky et al. (2018). The same is true for the reflectance of nonabsorbing snow layer. Therefore, we can derive both parameters in Eq. (1) from OLCI single view spectral reflectance measurements (see also Eq. (5))*.* Namely, it follows from dual-wavelength measurements (Kokhanovsky et al., 2018):

 (6)

(7)

The values of and are equal to the imaginary parts of ice refractive index at the wavelengths used for the determination of and *l* from OLCI reflectance measurements

and

An important point is the selection of OLCI channels used for the retrievals. These channels must satisfy to two conditions:

(1) The reflectance at these channels must be sensitive to the parameters of interest.

(2) These channels must be least influenced by atmospheric scattering and absorption processes.

The corresponding analysis shows that the channels located at 865 and 1020nm are the best candidates and we have used them in the retrieval process.

The derived value of *l* can be used to determine the snow spherical/plane albedo and also snow reflection function ( OLCI bottom of atmosphere reflectance) at any OLCI wavelength using Eqs. (1)-(5) and the imaginary part of ice refractive index at OLCI channels (see Appendix 2). The diameter *d* of ice grains in snow is estimated using the effective absorption length (Kokhanovsky, 2019):

(8)

where the parameter depends on the type of snow/shape of grains. We assume that *A=0.06* in the retrievalsas suggested by Kokhanovsky et al. (2019). The snow specific surface area is derived as

(8)

where is the bulk ice density. The derived spectral albedo is used to calculate the broadband albedo (BBA) using integration as shown below:

where is the incident solar flux at the snow surface, is plane (*p*) or spherical (*s*) albedo depending plane or spherical BBA is to be calculated. The indices 1 and 2 signify the wavelengths used. We have used the incident solar flux at the snow surface as derived from the code SBDART ( Ricchiazi et al., 1998) ( see Appendix 3). Generally, the results are only weakly sensitive to the variation of the function . Therefore, we have used use the same solar flux at the snow surface for the retrievals at different regions.

We have derived the following parameters with Eq. (9):

* visible spherical/plane BBA (),
* near IR spherical/plane BBA (),
* shortwave spherical/plane BBA ().

*In the case of clean snow*, Eq. (9) is applied directly because the spectral reflectance is known at the arbitrary wavelength (see Eqs. (1), (5)). The imaginary part of ice refractive index is taken from the study of Warren and Brandt (1994) (see Appendix 2). To speed up the retrieval process, we have used the approximation of the ratio given by Eq.(9) for the shortwave spherical albedo using the following equation:

, The coefficients are given in Table 3.

Table 3. The coefficients for the parametrization of the broadband spherical albedo in terms of the effective diameter of grains.

|  |  |  |  |  |
| --- | --- | --- | --- | --- |
| a | b | c |  | ,microns |
| 0.6420 | 0.1044 | 0.1773 | 158.62 | 2448.18 |

For the shortwave broadband plane albedo, we use the same equation as for the spherical albedo. However, the second order polynomial is used to represent the dependence of the coefficients *a,b,c,* , with respect to the cosine of the solar zenith angle (e.g., ). .

Table 4. The coefficients of the parametrization for the shortwave plane albedo.

|  |  |  |  |
| --- | --- | --- | --- |
|  |  |  |  |
| a | 0.7389 | -0.1783 | 0.0484 |
| b | 0.0853 | 0.0414 | -0.0127 |
| c | 0.1384 | 0.0762 | -0.0268 |
| 𝛥, microns | 187.89 | -69.2636 | 40.4821 |
| , microns | 2687.25 | -405.09 | 94.5 |

In case of polluted snow, we use the polynomial and exponential approximation for the snow albedo. This makes it possible to calculate BBA (spherical and plane) analytically. The details are given in Appendix 4.

**3.2 Retrievals for the polluted snow/ice case with account for atmospheric effects**

In case of polluted snow the simplified atmospheric correction based on the OLCI measurements at 865/1020nm wavelengths and extrapolation to the shorter wavelengths is not possible. This is due to the fact that impurities (and not just snow grains) influence the snow reflectance in the visible. Therefore, the atmospheric correction of measurements at all OLCI channels must be performed. Here, we consider two cases:

1) *the reflectance at 1020nm is larger than the THV assumed to be equal 0.5,*

2) *the reflectance at 1020nm is smaller or equal to the THV assumed to be equal 0.5.*

The first case corresponds to the polluted snow and the second one corresponds to other surfaces including bare ice. Lut us consider the first case now.

3.2.1 Polluted snow

In the case of polluted snow (R(1020nm)>0.5)we use the following equation to derive the snow spectral spherical albedo (Kokhanovsky et al., 2005):

(10)

is the measured top-of-atmosphere reflectance function, is atmospheric contribution to the measured signal, is the spherical albedo of the atmosphere, is the bottom-of-atmosphere snow reflectance, *T* is atmospheric transmittance from the top-of-atmosphere to the underlying surface and back to the satellite position, is the transmittance of purely gaseous atmosphere (see Appendix 5). Eq. (10) is valid for Lambertian surfaces with . The snow is not exactly Lambertian reflector, therefore, we have assumed that (see Eq. (1) ) in the nominator of Eq. (10). The value of is derived in the same way as for clean snow at OLCI reflectances above 0.5. Otherwise, the analytical approximation ( Kokhanovsky et al., 2019) is used.

As in the case of clean snow, we shall assume that scattering and absorption of light by atmosphere and impurities in snowpack can be ignored at the wavelengths 865 and 1020nm. This makes it possible to derive the parameters , *l, d, as described above for the case of clean snow.* However, Eq. (5) can not be used because the spectral reflectance in the visible is influenced not just by ice grains but also by various impurities (soot, dust, algae). The value of can be derived analytically from Eq. (10) under assumption that

parameter from the transcendent Eq. (10) for the pre-described atmospheric aerosol properties with account for gaseous absorption due to ozone and water vapor (the terms , , and *T,*  see Appendix 5) at all channels except those affected by the molecular absorption by oxygen (channels 13-15). The spherical albedo at these channels is found assuming the linear spectral behavior of spherical albedo in the range 753-778nm and using OLCI measurements at 753 and 778nm. We approximate the gaseous transmittance term as follows:

(11a)

where *m* is the air mass factor, is the vertical optical depth of ozone at the concentration

*c=405DU*,

. (11b)

Here, is the ozone concentration provided in the OLCI satellite file (with account for units). In particular, to transfer from OLCI O3 units (kg/) to Dobson Units (DU), we multiply OLCI O3 concentration by a constant factor equal to 4.6729e+4. Therefore, the total ozone load 300DU corresponds to 6.42e-3 kg/. The values of calculated for all OLCI channels with account for the instrument response function are given in Appendix 4. We approximate air mass factor as follows:

. (11c)

At the wavelengths larger than 865nm, the influence of pollutants is small and we use the derived EAL and Eq.(5) to find the spectral spherical albedo. The plane albedo is derived using Eq. (4) and the bottom-of-atmosphere snow reflectance is derived using Eq. (1). The BBA is found from Eq.(9).

The concentration of pollutants in snow is estimated using the approach described below. It is assumed that the spherical albedo can be modelled using the modified Eq. (5):

, (12)

where is spectral absorption coefficient of impurities. The parameter can be derived using the following approximation (Kokhanovsky et al., 2018):

, (13)

where *B* is so-called absorption enhancement parameter for ice grains (Kokhanovsky et al., 2019) and is the volumetric concentration of ice grains in snowpack. We shall assume that *B=1.6* in the retrieval procedure. Eq. (12) can be used to derive the normalized absorption coefficient of impurities . Namely, it follows from Eq. (12):

, (14)

Where is given by Eq. (11). The relative volumetric concentration of pollutants in snow can be derived from the measurements at the wavelength 400nm. Namely, it follows from Eq. (14):

, (15)

where we ignored absorption by ice particles, which is valid approximation in the visible, and accounted for the fact that , is the volumetric absorption coefficient of impurities. Eq. (15) shows that the determination of the concentration of pollutants is possible only if the volumetric absorption coefficient of pollutants is known in advance. In particular, it can be approximated as(Kokhanovsky et al., 2018)

(16)

for soot. Here, is the bulk absorption coefficient of soot , is the imaginary part of soot refractive index, and *p=0.9* (Kokhanovsky et al., 2018). We provide the concentrations of pollutants using Eq. (15) and assuming that in the output of the algorithm, if the algorithm identifies soot as an impurity in snow. If the dust is assumed to be a major impurity in snow, we assume: One needs to scale the results using a particular value of the volumetric absorption coefficient of pollutants (at 400nm) for a given site. The determination of absorption Angström exponent for impurities is discussed in Appendix 6.

3.2.2 Bare ice and other surfaces

*Spectral snow characteristics*

If the underlying surface is not snow, the application of the equation relating the snow albedo to the snow grain size is not justified. In this case the spherical albedo is found for all OLCI channels (except 19 and 20) using Eq. (10). It is assumed that the value of reflectance for a nonabsorbing surface can be approximated as discussed by Kokhanovsky et al. (2019) (see Appendix 7). This enables also the calculation of the snow planar albedo and also the snow reflectance. For the determination of the snow albedo and reflectance at channels affected by the absorption by water vapor (19 and 20), the linear interpolation in the range 885-1020nm is used.

*Broadband albedo*

For the calculation of the BBA, one must have the functional dependence of the spherical albedo at arbitrary wavelength. It is derived using the following assumptions:

1. the spherical albedo can be approximated by the polynomial of the second order on the interval 400-709nm; the corresponding coefficients are derived from the values of reflectance at 400, 560, and 709nm.
2. the spherical albedo can be approximated by the polynomial of the second order on the interval 709-865nm; the corresponding coefficients are derived from the values of reflectance at 709, 753, and 865nm.
3. the spherical albedo can be approximated by the exponential function at the wavelengths above 865nm; the corresponding coefficients are derived from the values of reflectance at 865 and 1020nm.

These assumptions make it possible to derive the value of BBA analytically (see Appendix 4).

**4. Flags**

Several flags are introduced in the snow processor. They are explained in this section.

The snow flag is determined by the value of OLCI normalized difference snow index (NDSI):

(17)

The snow flag is equal to one (100% snow – covered pixel), if NDSI is in the range 0.3-0.4 and R(400nm) is larger than 0.75.

The bare ice flag is determined by the value of OLCI normalized difference bare ice index (NDBI):

(18)

The bare ice is classified in two steps. First, dark bare ice is identified where NDBI is less than 0.65 and R (400nm) is less than 0.75. Then for cases the dark bare ice flag is not set, the bare ice flag is equal to one (100% bare ice – covered pixel), if NDSI is larger than 0.33. Also is assumed that the dark dirty bare ice flag is equal to one (100% dark dirty bare ice – covered pixel), if NDBI is smaller than 0.65 and R (400nm) is smaller than 0.75 and that a land mask is used.

The values of NDSI and NDBI are provided in the output of the algorithm. In principle, the value of NDSI can be used for the estimation of snow fraction in the OLCI pixel.

**Appendix 1. Definitions: reflectance, spherical and plane albedos**

The top-of-atmosphere reflection function is defined as (Kokhanovsky, 2006)

(A1.1)

Here, is the intensity of reflected light, is the solar flux at the top-of-atmosphere, is the cosine of the solar incidence angle . Many satellite instruments simultaneously measure both and . Therefore, the reflection function can be easily found using Eq. (A1.1) as measured by satellite sensors. We shall consider only cloud free pixels in this work. The reflection function depends on atmospheric effects due to molecular and aerosol scattering and absorption of solar radiation. For retrieval of surface optical properties, these effects must be removed. Then the bottom-of atmosphere reflectance function *R* is determined and used to derive the *plane albedo* as:

(A1.2)

where

(A1.3)

is the relative azimuthal angle, is the cosine of the viewing zenith angle . The *spherical albedo* is found by integration of the spherical albedo:

(A1.4)

While the procedure is straightforward, space-borne optical instruments usually perform single angle spectral observations and do not sample all observation angles for a given pixel. This problem is solved using multiple observations of the same target during several days sampling observation geometries and finding parameters of a prescribed surface reflectance model.

**Appendix 2. The spectrum of imaginary part of ice refractive index**

**Table A2.1. The imaginary part of ice refractive index at OLCI channels.**

|  |  |  |  |  |  |  |  |  |  |  |  |
| --- | --- | --- | --- | --- | --- | --- | --- | --- | --- | --- | --- |
|  | 400 | 412 | 442 | 490 | 510 | 560 | 620 | 665 | 673 | 681 | 708 |
|  | 2.365e-11 | 2.7e-11 | 7.0e-11 | 4.17e-10 | 8.04e-10 | 2.84e-9 | 8.58e-9 | 1.78e-8 | 1.95e-8 | 2.10e-8 | 3.3e-8 |
|  | 753 | 761 | 764 | 767 | 778 | 865 | 885 | 900 | 940 | 1020 |  |
|  | 6.23e-8 | 7.1e-8 | 7.68e-8 | 8.13e-8 | 9.88e-8 | 2.4e-7 | 3.64e-7 | 4.20e-7 | 5.53e-7 | 2.25e-6 |  |

**Appendix 3. The spectrum of incident solar flux at the bottom of atmosphere**

The spectrum of incident radiation at the snow surface is approximated by the following analytical equation:

, (A3.1)

where the coefficients are presented in Table A3.1. The coefficients have been derived using

the code SBDART(Ricchiazi et al., 1998) in the spectral range 300-2400nm at the following assumptions:

* the water vapour column: 2.085g/,
* the ozone column: 0.35 atm-cm,
* the tropospheric ozone: 0.0346atm-cm,
* the aerosol model: rural (Shettle and Fen, 1979),
* the vertical optical depth of boundary layer at 550nm: 0.1,
* the altitude: 825m,
* the solar zenith angle 60 degrees,
* the snow albedo at the surface: calculated using the assumption that the grain diameter is equal to 0.25mm and grains are of spherical shape

**Table A3.1. The coefficients of approximation**

|  |  |  |  |  |
| --- | --- | --- | --- | --- |
|  |  |  | , microns | , microns |
| 3.238e+1 | -1.6014033e+5 | 7.95953e+3 | 8.534e-4 | 4.0179e-1 |

**Appendix 4. The broadband albedo of polluted snow and ice**

The broadband albedo can be found using the following equation:

where

(A4.2)

and the spectral albedo is modelled as follows:

(A4.3)

for the OLCI spectral range and as

(A4.4)

at the wavelengths larger than 1020nm. Integral (A4.1) with account for Eqs. (A4.2), (A4.3) can be evaluated analytically. The answer is:

*cL,* (A4.5)

where

,

,

.

Here is the integral given in the dominator in Eq. (A4.1) (evaluated analytically with account for Eq. (A4.2)) and

,,,

,

,

.

It follows with account for Eqs. (A4.1), (A.4.4):

*,*

where

*)*

is the analytical expression for the nominator of Eq. (A4.1) for the wavelengths larger than 1020nm.

**Appendix 5. The simulation of top-of-atmosphere reflectance with account for molecular and aerosol light scattering effects**

The background atmospheric aerosol in Arctic is usually characterized by the low values of aerosol optical thickness and values of single scattering albedo close to one. Therefore, one can neglect light absorption by aerosol and assume that the atmosphere-underlying surface reflectance (due to molecular and aerosol scattering and reflectance from underlying surface) can be presented in the following way:

, (A5. 1)

where is the spherical albedo of underlying surface, is spherical albedo of atmosphere, is the total atmospheric transmittance from the top-of-atmosphere to the underlying surface and from there to an optical instrument on board of a satellite. The snow reflectance function in Eq. (4.1) is related to the snow spherical albedo by means Eq.(1).

The atmospheric reflectance due to coupled aerosol-molecular scattering can be presented in the following way (in the framework of the Sobolev approximation (Sobolev, 1975):

(A5.2)

where single scattering contribution

(A5.3)

and multiple light scattering contribution is approximated as

(A5.4)

where

,, (A5.5)

,, (A5.6)

(A5.7)

Here, is the cosine of the solar zenith angle (SZA), is the cosine of the viewing zenith angle (VZA) , is the scattering angle defined as

, (A5.8)

is the relative azimuthal angle (equal to 180 degrees minus OLCI relative azimuthal angle), is the sine of the SZA, is the sine of the VZA, is the atmospheric optical thickness, is the phase function, is the asymmetry parameter defined as

. (A5.9)

The transmission function is approximated as follows:

, (A5.10)

where is calculated using the following approximation:

, (A5.11)

(A5.12)

is the so – called backscattering fraction. The spherical albedo is found using the approximation proposed by Kokhanovsky et al. (2007):

. (A5.13)

The coefficients of polynomial expansions of all coefficients (a,b,c, ) in Eq.(4. 13) with respect to the value of *g* are given by Kokhanovsky et al. (2005).

One can see that the reflection function depends on the atmospheric optical thickness, which can be presented in the following form:

. (A5.14)

The molecular optical thickness can be approximated as

, (A5.15)

where , *p* is the site pressure, and the wavelength is in microns. We calculate the site pressure using the following equation: *z* is the height of the underlying surface provided in OLCI files and *H=7.64km* is the scale height.

It follows for the aerosol optical thickness (AOT):

, (A5.16)

where . The pair represents the Angström parameters. Currently, we use the fixed values of Due to low aerosol load in Arctic, this assumption does not lead to the substantial errors.

The phase function can be presented in the following form:

, (A5.17)

where

(A5.18)

is the molecular scattering phase function and is the aerosol phase function. We shall represent this function as:

. (A5.19)

Therefore, it follows for the asymmetry parameter:

. (A5.20)

The parameter varies with the location, time, aerosol, type, etc. We shall assume that it can be approximated by the following equation:

. (A5.21)

The coefficients in this equation ( as derived from multiple year AERONET observations over Greenland) are as follows:

(A5.22)

The parameter *B* for the phase function (A5.17) has the following form (see Eq. (A5.12)):

, (A5.23)

where =0.5 and (Kokhanovsky et al., 2020).

It should be pointed that the system of equations given above enables the callculation of underlying snow-atmosphere reflectance as a function of the aerosol optical thickness for a known value of the snow spherical albedo.

The gaseous transmittance term can be found using the following analytical approximation:

(A5.24)

where is the vertical optical depth of ozone at the concentration *c=405DU*,

. (A5.25)

Here, is the ozone concentration provided in the OLCI satellite file (with account for units). In particular, to transfer from OLCI O3 units (kg/) to Dobson Units (DU), we multiply OLCI O3 concentration by a constant factor equal to 4.6729e+4. Therefore, the total ozone load 300DU corresponds to 6.42e-3 kg/. The values of calculated for all OLCI channels with account for the instrument response function are given in Table A5.1.

Table A5.1. The spectral dependence of ozone vertical optical thickness in terrestrial atmosphere at the ozone load equal to 405 DU.

400.00000 1.378170469E-004

412.50000 3.048780958E-004

442.50000 1.645714060E-003

490.00000 8.935947110E-003

510.00000 1.750535146E-002

560.00000 4.347104369E-002

620.00000 4.487130794E-002

665.00000 2.101591797E-002

673.75000 1.716230955E-002

681.25000 1.466298300E-002

708.75000 7.983028470E-003

753.75000 3.879744653E-003

761.25000 2.923775641E-003

764.37500 2.792211429E-003

767.50000 2.729651478E-003

778.75000 3.255969698E-003

865.00000 8.956858078E-004

885.00000 5.188799343E-004

900.00000 6.715773241E-004

940.00000 3.127781417E-004

1020.00000 1.408798425E-005

**Appendix 6. Angström absorption parameter for impurities in snow**

The polluted snow spherical albedo can be presented in the following form in the visible:

. (A6.1)

Let us assume that the impurity absorption coefficient has the following form:

, (A6.2)

where Then one can easily derive from equation (6.1) for the Angström absorption coefficient:

, (A6.3)

where

. (A6.4)

Here are spherical albedos at the wavelengths It follows for the value of

*/l .* (A6.5)

We assume that the snow is polluted by black carbon if *m* is smaller than 1.2. Otherwise, the dust pollution is assumed.

**Appendix 7. The approximation for the reflectance of nonabsorbing snow**

It is assumed that the reflection function of a semi-infinite snow layer can be approximated by the following expression:

|  |
| --- |
|  |

(A7.1)

where *A* = 1.247, *B* = 1.186, *C* = 5.157,

|  |  |
| --- | --- |
|  |  |

and θ is the scattering angle in degrees.

It holds:

(A7.2)

|  |  |
| --- | --- |
|  |  |

Here, *s* and are the sines of the viewing zenith angle (VZA) and SZA, respectively, *φ* is the relative azimuthal angle and (, ) are the cosines of the SZA and VZA, respectively. The OLCI relative azimuth angle υ must be transformed as: *φ* = abs(180 − υ) to be used in the equations given above.

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