**-SMOS Level 3 SSS Research products - Algorithm Theoretical Breadboard Document**

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1. Introduction

SMOS (Soil Moisture and Ocean Salinity) mission is a joint ESA / CNES / CDTI Earth Observation program. The SMOS mission proposed by CESBIO has been selected as the 2nd Earth Explorer Opportunity Mission. The SMOS satellite was launched the November 2, 2009.

The objective of the SMOS mission is to provide Soil Moisture (SM) and Ocean Salinity (OS) maps. Both SM and OS are key variables in climate monitoring, surface / vegetation / atmosphere transfers, and ocean / atmosphere cycles.

The SMOS Satellite Control Centre has been adapted by CNES from generic PROTEUS ground segment and installed in Toulouse within the CNES premises. A SMOS Mission Centre dedicated to level 1 and 2 products, has been developed and located in the ESA / Villafranca centre (Data Processing Ground Segment). A specific Data Processing Centre dedicated to level 3 and 4 products has been developed by CNES (CATDS).

The Centre Aval de Traitement des Données SMOS (CATDS), developed by the CNES in collaboration with the CESBIO and IFREMER, is dedicated to:

- Product and distribute SMOS L3/L4 products (from L1B products from the ESA Data Processing Ground Segment)
- Reprocess SMOS L3/L4 products when necessary
- Propose improvements of the L0 to L2 processing chains (possibly helping to fine tune the calibration aspects) ; once validated, these algorithms will be proposed for a transfer to ESA Data Processing Ground Segment.
- Provide services & hot-line support to L3/L4 users
- Develop, test and validate algorithms for L3 and L4 processing chains, in close cooperation with the scientific community

The CATDS is made up of 3 centers:

- Data Production Center (CPDC), host by IFREMER Brest
- Soil Moisture Expertise Center (CEC-SM), host by CESBIO Toulouse
- Ocean Salinity Expertise Center (CEC-OS), host by IFREMER Brest

The objective of this Algorithm Theoretical Breadboard Document (ATBD) is to describe the first version of the Algorithm used to generate the CATDS/CEC-OS Sea Surface Salinity Level 3 composite products.
1.1 SMOS product Definition

SMOS DATA PROCESSING FLOW

The SMOS data processing flow is illustrated and explained in the diagram below. Click on data for explanations.

1.1.1 Level 0

These are SMOS Payload data in so-called Source Packets with added Earth Explorer product headers. They are chronologically sorted by Source Packet type: Observation Data and Housekeeping Telemetry.

1.1.2 Level 1A: Calibrated Visibilities

These are the SMOS reformatted and calibrated Observation and Housekeeping data in engineering units. Level 1A products are physically consolidated in pole-to-pole time-based segments. Scientific SMOS Level 1A products are the so-called "Calibrated Visibilities".
1.3 Level 1B:

The SMOS Level 1B products are the output of the image reconstruction of the SMOS observation measurements and consist of geo-located vectors of Brightness Temperatures in the antenna polarisation reference frame.

1.4 Level 1C

Since Level 1B products are arranged as snapshots and not geographically sorted, SMOS Level 1C products constitute reprocessed Level 1B, which are geographically sorted, that is swath-based maps of Brightness Temperature.

1.5 Level 2

Level 2 OS products are geo-located geophysical (Sea Surface Salinity) products with all the calibration applied and with neither spatial nor temporal averaging. An inversion algorithm is applied to the set of the brightness temperatures from L1C. L2 OS products keep the resolution of L1C. Level 2 products are Ocean Salinity swath-based maps.

1.6 Level 3 Composite

L3 corresponds to re-sampled and temporally accumulated Sea surface Salinity data. At the CATDS/CEC-OS, the L3 are computed from L1A data via a simplified & bug-corrected image reconstruction algorithm, spatial and temporal aggregation of brightness temperature and improvement for taking into account the instrument errors (calibration) and the data spatio-temporal variabilities (geophysical, RFI). L3 is based on more auxiliary data (and more consolidated auxiliary data) than L2 data as it has less operational constraints.

1.2 CATDS/CEC-OS "research" Level 3 SSS products characteristics

Based on up-to-date algorithms using the ESA reprocessed L1A data for year 2010 as inputs, the CATDS-CEC OS as produced one year of composite Level 3 SSS data.

These so-called “CATDS/CEC-OS SMOS Level 3 SSS research products” are of three types depending on the temporal window used to generate the Level 3 sea surface salinity composite.

The temporal windows we used to generate the products are:

- monthly composite
- 10 days composite
- daily composite based on a running +/-5 days temporal window

For each temporal resolution, the composite products are gridded at several spatial resolution varying from 0.25° to 1°, depending on the temporal window:

- Monthly composite products are available at a spatial resolution of:
  - 0.25° x 0.25°
0.50° x 0.50°
1.00° x 1.00°

- 10 days composite products are available at a spatial resolution of:
  0.25° x 0.25°
  0.50° x 0.50°
  1.00° x 1.00°

- Daily composite products based on a running +/-5 days temporal window are for the moment only available at 0.25° x 0.25°

  The products generated are in Netcdf format and can be accessed at the new CATDS web site http://www.catds.fr/
  under =>Data=>Research Products from Expertise Centers
  The data access is under an ftp server:

  ftp : eftp.ifremer.fr, password protected
  The access is free upon an email request: support@catds.fr

  The data are provided with an associated documentation including a Product User Manual, a Validation Report and the present ATBD.
2. Algorithm Overview

2.1 From Calibrated Visibilities to Multi-angular daily surface emissivities

- ESA reprocessed L1A
- JRECON image Reconstruction algorithm
- First Stokes TBs at antenna Level
- Fixed OTT Corr
- Atmospheric Galactic noise corrections
- First Stokes TBs at surface Level AFOW
- Binning in 10 incidence angle bins 5° width 15°-60°
- Down sampling & gridding @ 25 km res Daily Asc/desc separately
- Daily Gridded First Stokes Surface Emissivities $e^{\text{SSS}_{\text{vis}}}(\text{lat,lon}, \theta_{\text{inc}}=15^\circ-60^\circ, \text{doy}=1-365)$
- $e^{\text{SSS}_{\text{mic}}}(\text{lat,lon}, \theta_{\text{inc}}=15^\circ-60^\circ, \text{doy}=1-365)$
- Co-located ECMWF SST data
3. **Input Data & reprocessed L1 analysis**

SMOS L1A (calibrated visibilities) and L1B (brightness temperatures) data acquired during the period January 2010 to December 2010 have been reprocessed by ESA in February 2010. The intention of this intermediate reprocessing was to provide a consistent data set for the first year of mission operations, since there were several updates of the SMOS instrument settings and processors throughout the commissioning phase (which is normal procedure for any commissioning phase). The data has been processed with the prototype algorithm baseline as available at the end of commissioning phase (V346) and with the following assumptions for calibration activities:

- Fixed calibration baseline for what concern the Flat Target Correction FTT (acquired on 10 July 2010), the Noise Injection Radiometer (NIR) calibrated on 14 July 2010 and the instrument response function (G/J matrix) calibrated on 2 February 2010.

- Long-Calibration to update the PMS gain and offset and the visibility offset was performed according to routine calibration plan.

- Local-Oscillator calibration to update the phase of the fringe washing Function at the origin FWF(0) was performed according to routine calibration plan.

The above fixed calibration parameters were used in the reprocessing to tentatively better understand the sources for the several problems encountered in SMOS Level 2 products, and in particular, for the measurements over the oceans.
4. **JRECON Image Reconstruction algorithm**
5. From Reconstructed TB images to daily surface emissivities

5.1. Ocean Target Transformation

5.2. Corrections for Atmospheric contributions

Our forward model algorithm for atmospheric corrections at L-band differs from the ESA operational Level 2 algorithm. It is therefore described in detail in what follows. Main differences with the ESA processor are listed here below:

- Auxiliary atmospheric parameters (pressure, temperature, relative humidity, cloud water,..) are obtained from NCEP GFS 6-hourly fields at multiple altitudes above ground

- Attenuation rates and atmospheric absorption are estimated on N-layers along the vertical from ground to satellite altitude using the Milimeter Wave Propagation Model and finally integrated along the path from target to satellite.

ESA operational Level 2 algorithm uses a vertically integrated formulation and rely on ECMWF parameters.

5.2.1. Electromagnetic Wave Propagation through the Atmosphere

We will assume that, at L-band, the attenuation of radiation propagating through the atmosphere is mostly associated with the electric field components in the plane normal to the propagation direction.

\[
\begin{align*}
E_h(r,t) &= \Re \left\{ \hat{E}_h(r,t) \right\} = \Re \left\{ a_h e^{i(kr-\omega t-\phi_h)} \right\}, \\
E_v(r,t) &= \Re \left\{ \hat{E}_v(r,t) \right\} = \Re \left\{ a_v e^{i(kr-\omega t-\phi_v)} \right\},
\end{align*}
\]

where \(\phi_h\) and \(\phi_v\) are constant phases, \(r\) is distance along the propagation path, \(k\) is wavenumber magnitude, \(\omega\) is the radiation frequency, \(t\) is time, and the real amplitudes \(a_h(r)\) and \(a_v(r)\) are functions of propagation distance \(r\). In general as the radiation propagates through the atmosphere it experiences changes in the dielectric properties of the medium. We will assume that the amplitude coefficients satisfy the simple ordinary differential equations:

\[
\begin{align*}
\frac{da_h}{dr} &= -b(r)a_h(r), \\
\frac{da_v}{dr} &= -b(r)a_v(r).
\end{align*}
\]
In reality, the amplitude attenuation rate $b(r)$ is a function of space and time and may be computed using molecular oxygen ($O_2$) and water vapor concentration profiles along with atmospheric physical temperature and pressure profiles. The attenuation rate, or attenuation coefficient, $b$ (typically expressed in units of nepers per meter), generally is a complicated function of atmospheric state (i.e. pressure, temperature, water content in various forms, etc.). Multiplying (5.2.1) by $a_h(r)$ and (5.2.2) by $a_v(r)$ and rearranging yields:

$$\frac{da_h^2}{dr} = -2b(r)a_h^2(r), \quad (5.2.3)$$

$$\frac{da_v^2}{dr} = -2b(r)a_v^2(r). \quad (5.2.4)$$

Similarly, multiplying (5.2.1) by $a_v(r)$ and (5.2.2) by $a_h(r)$ and adding yields

$$\frac{da_v a_h}{dr} = -2b(r)a_v a_h(r). \quad (5.2.5)$$

If the attenuation rate $b$ is independent of spatial location and time, and if we let the electric field amplitudes at some reference location (at which $r = 0$) be $a_{vo}$ and $a_{ho}$, then

$$a_h(r) = a_{ho}e^{-br}, \quad a_v(r) = a_{vo}e^{-br},$$

so that

$$E_h(r,t) = \Re\left\{ \hat{E}_h(r,t) \right\} = \Re\left\{ a_{ho}e^{i(kr-\omega t-\phi_h)-br} \right\},$$

$$E_v(r,t) = \Re\left\{ \hat{E}_v(r,t) \right\} = \Re\left\{ a_{vo}e^{i(kr-\omega t-\phi_v)-br} \right\},$$

where $a_{ho}$ and $a_{vo}$ are the real amplitude of the components at some reference location $r=0$. We may also represent this attenuation rate as an imaginary part of the wavenumber magnitude $k$, so that we let:

$$\tilde{k} = k + ib = k_r + ik_i,$$

where now $\Re(\tilde{k})$ gives the propagation speed in terms of $\omega$ while $\Im(\tilde{k})$ gives the attenuation rate.
The complex wavenumber magnitude $\tilde{k}$ is the key quantity required to model the radiation propagation (including curvature in propagation path) and attenuation along the path, and here we use the Millimeter-wave Propagation model of [Liebe et al., 1993] (hereafter referred to as the MPM93), which is a refined version of the model presented in [Liebe et al., 1989], to compute $\tilde{k}$ as a function of atmospheric state along the radiation propagation path. As $\tilde{k}$ depends mostly on molecular oxygen concentration and, to a lesser extent, water vapor concentration, the key atmospheric variables required are dry air pressure and water vapor mixing ratio.

The Millimeter-wave Propagation model provides an explicit expression for the so-called complex index of refraction, denoted by $N$, which is defined so that

$$\tilde{k} = \tilde{k}_r + i\tilde{k}_i = k + i\frac{\tilde{N}}{k} = k + k(N_0 + N') + iN'',$$

where $N_0$, $N'$, and $N''$ are all real quantities that depend upon the atmospheric state. $N_0$ is independent of frequency while $N'$ and $N''$ depend on frequency, in general. The real and imaginary parts of complex index of refraction are related to the real and imaginary parts of the complex wavenumber by the equations

$$\tilde{k}_r = k(1 + N_0 + N'),$$
$$\tilde{k}_i = b(r) = kN''.$$

Thus, $N''$ corresponds to the loss portion of $\tilde{k}$ while $N_0 + N'$ corresponds to the propagation speed portion of $\tilde{k}$. It follows that the corresponding amplitude attenuation rate, $b(r)$, expressed in nepers per meter, is

$$b(r) = kN'' = \left(\frac{2\pi\nu}{c}\right)N'' \quad [\text{Np m}^{-1}],$$

and the power attenuation rate is

$$a(r) = 2b(r) = 2kN'' = \left(\frac{4\pi\nu}{c}\right)N'' \quad [\text{Np m}^{-1}],$$

where $c$ is the speed of light in a vacuum (m s$^{-1}$) and $\nu$ is the electromagnetic frequency in hertz. The index of refraction as provided by the MPM93, $\bar{N} = N_0 + N' + N''$ is expressed in ppm (i.e., it is scaled by $10^6$) so that for $b(r)$ in nepers per meter, we have
\begin{equation}
\tilde{k}_r(r) = b(r) = kN'' = \left(\frac{2\pi\nu}{c}\right) \cdot 10^{-6} \cdot \tilde{N}'' \quad \text{[Np m}^{-1}\text{]},
\end{equation}

or for \( b(r) \) in dB km\(^{-1}\) with \( \nu \) in gigahertz,
\begin{equation}
b(r) = kN'' = \left(\frac{2\pi\nu}{c}\right) \left(\frac{10}{\log_{10}(10)} \text{ dB Np}\right) \left(10^3 \text{ m km}\right) \left(10^6 \text{ Hz GHz}\right) \cdot 10^{-6} \cdot \tilde{N}'' \approx 0.0010\nu N'' \quad \text{[dB km}\^{-1}\text{].}
\end{equation}

The real part of the complex wavenumber, \( \tilde{k}_r(r) \), is
\begin{equation}
\tilde{k}_r(r) = \frac{360}{2\pi} \left(\frac{\text{deg}}{\text{rad}}\right) \left(\frac{2\pi\nu}{c}\right) \left(10^6 \text{ Hz GHz}\right) \left(10^3 \text{ m km}\right) \left[1 + 10^{-6} \cdot \left(\tilde{N}_0 + \tilde{N}'\right)\right] \quad \text{[deg km}\^{-1}\text{].}
\end{equation}

The variable part of the preceding expression for \( \tilde{k}_r(r) \) is called the *phase dispersion*, \( \beta \), and is approximately
\begin{equation}
\beta = 1.2\nu \left(\tilde{N}_0 + \tilde{N}'\right) \quad \text{[deg km}^{-1}\text{]},
\end{equation}

with \( \nu \) expressed in gigahertz and \( \tilde{N}_0 \), and \( \tilde{N}' \) expressed in ppm. From \( \tilde{k}_r(r) \) we can derive the wave propagation speed \( \tilde{c}(r) \):
\begin{equation}
\tilde{c}(r) = 1000 \cdot \frac{\omega}{\tilde{k}_r(r)} \left(\frac{\text{deg s}^{-1}}{\text{deg km}^{-1}}\right) = 1000 \cdot \frac{360\nu}{\tilde{k}_r(r)} \left(\frac{\text{deg s}^{-1}}{\text{deg km}^{-1}}\right) = c \left[1 + 10^{-6} \cdot \left(\tilde{N}_0 + \tilde{N}'\right)\right]^{-1} \quad \text{[m s}^{-1}\text{].}
\end{equation}

where as before \( c \) is expressed in meters per second and \( \sim N_0 \) and \( \sim N_0' \) are expressed in ppm as in the MPM93. If we were concerned about propagation path curvature, we would need to use this expression to derive the propagation path for each point in director cosine coordinates. But we will assume that at L-band such curvature effects are negligible. At L-band, the attenuation rate \( b(r) \) is associated mostly with the long tails in the distributions of absorption associated with resonant absorption lines of molecular oxygen and water vapor at much higher frequencies. Scattering by cloud droplets, ice, and raindrops is relatively minor at L-band.

Next we will examine the formulation of \( \tilde{N}_0 \), and \( \tilde{N}' \) in the MPM93 and then relate the total refractive index \( \tilde{N}_0 \), and \( \tilde{N} \) to the Stokes vector, which is our primary concern.

### 5.2.2. Formulation of the Atmospheric Refractive Index

As formulated in the MPM93, the total index of refraction is the sum of indices of refraction owing to molecular oxygen (\( \tilde{N}_0 \)), water vapor (\( \tilde{N}_v \)), and cloud water (\( \tilde{N}_w \)) and ice (\( \tilde{N}_i \)), so that:
\[ N(T_r, e, p_d, \nu) = \tilde{N}_0 + \tilde{N}^l + i\tilde{N}^m = \tilde{N}_D(T_r, e, p_d, \nu) + \tilde{N}_Y(T_r, e, p_d, \nu) + \tilde{N}_W(T_r, \rho_w, \nu) + \tilde{N}_I(T_r, \rho_i, \nu) \]

where \( p_d \) and \( e \) are the partial pressures of dry air and water vapor (hPa), respectively; \( \rho_w \) and \( \rho_i \) are the liquid water ice densities (g m\(^{-3}\)), respectively; \( T_r \) is the reciprocal temperature, defined by

\[ T_r = \frac{300}{T_p}, \]

where \( T_p \) is the physical temperature in kelvin. The indices of refraction are expressed in the MPM93 in terms of the reciprocal temperature, partial pressures of dry air and water vapor (which is a function of relative humidity and temperature), and cloud water and ice densities (not mixing ratios). The molecular oxygen concentration is a function of the partial pressure of dry air, \( p_d \) (expressed in hPa), which is approximately equal to the total pressure \( p_t \) minus the partial pressure of water vapor \( e \):

\[ p_d = p_t - e, \]

Following Liebe et al., 1993, we introduce the following approximation to the water saturation vapor pressure over a plane surface of water:

\[ e_s(T_r) = 2.408 \times 10^{11} T_r^5 e^{-22.644 T_r}, \]

Using this approximation, the partial pressure of water vapor in hectopascals is

\[ e(T_r, r) = e_s(T_r) \frac{r}{100} = \frac{r}{100} \cdot 2.408 \times 10^{11} T_r^5 e^{-22.644 T_r}, \]

where \( r \) is the relative humidity in percent.

The dry air index of refraction owing to molecular oxygen \( (\tilde{N}_o) \), consists of a nondispersive term \( \tilde{N}_d \), a nonresonant term \( \tilde{N}_n \), and a sum over relevant molecular oxygen absorption lines:

\[ \tilde{N}_D(T_r, e, p_d, \nu) = \tilde{N}_d + \tilde{N}_n + \sum_{k=1}^{44} S_k F_k \text{ [ppm]}. \]

The nondispersive term is

\[ \tilde{N}_d = 0.2588 p_d T_r \text{ [ppm]}, \]

and the nonresonant term is given by the sum of a relaxation spectrum term \( S_o F_o \) and a term associated with pressure-induced \( N_o \) absorption (which makes a small contribution above 100 GHz), \( S_n F_n \):

\[ \tilde{N}_n = S_o F_o + iS_n F_n \text{ [ppm]}, \]

with
In the sum over absorption lines, \( \sum_{k=1}^{44} S_k F_k \), the individual line strengths are given by

\[
S_k = 1 \times 10^{-6} \frac{a_1}{\nu_k} p_d T_r^3 \exp \left[ a_2 (1 - T_r) \right] \quad \text{[ppm]},
\]

where \( a_1 \) and \( a_2 \) are constants for each line. The line form (or line spreading) factor, known as the Van Vleck-Weisskopf function, is given by

\[
F_k = \nu \left[ \frac{1 - i \delta_k}{\nu_k - \nu - i \gamma_k} - \frac{1 + i \delta_k}{\nu_k + \nu + i \gamma_k} \right],
\]

with the overlap parameter \( \delta_k \) given by

\[
\delta_k = 1 \times 10^{-3} (a_5 + a_6 T_r) (p_d + e) T_r^{0.8},
\]

where \( a_5 \) and \( a_6 \) are constants for each line. The width parameter \( \gamma_k \) is given by

\[
\gamma_k = \left( \frac{\tilde{\gamma}_k^2 + 2.25 \times 10^{-6}}{2} \right)^{1/2} \quad \text{[GHz]},
\]

with

\[
\tilde{\gamma}_k = 1 \times 10^{-3} a_3 (p_d T_r^{1.4} + 1.1 e T_r) \quad \text{[GHz]},
\]

where \( a_3 \) and \( a_d \) are constants for each line. The difference between \( \gamma_k \) and \( \tilde{\gamma}_k \) is related to the Zeeman effect in which absorption lines are split into multiple lines.

The index of refraction for water vapor is the sum of a nondispersive term, a continuum term, and a sum over 35 resonant absorption lines (including one pseudo-line at 1780 GHz). At L-band the continuum term is negligible so that

\[
\tilde{N}_v \approx \tilde{N}_r + \sum_{k=1}^{35} S_k^{(v)} F_k^{(v)} \quad \text{[ppm]},
\]

The nondispersive term is

\[
\tilde{N}_r(T_r, e, p_d, \nu) = (4.163 T_r + 0.239) T_r \quad \text{[ppm]},
\]

The absorption line strength in the sum over the lines is

\[
S_k^{(v)} = \left( \frac{b_1}{\nu_k} \right) e T_r^{3.5} \exp \left[ b_2 (1 - T_r) \right],
\]
where \( b_1 \) and \( b_2 \) are constants defined for each line \( k \). The line form function has the same form as for oxygen except that there is no line overlap effect:

\[
F_k^{(v)} = \nu \left[ \frac{1}{\nu_k^{(v)} - \nu - i\gamma_k^{(v)}} - \frac{1}{\nu_k^{(v)} + \nu + i\gamma_k^{(v)}} \right].
\]

For a total pressure above 0.7 hPa, the width of a pressure-broadened vapor line, \( \gamma_k^{(v)} \) is given by

\[
\gamma_k^{(v)} = 1 \times 10^{-3} b_3 (b_4 e(\epsilon_v(T_r), r)T_k^{66} + p_kT_k^{56}) \quad \text{[GHz]},
\]

where \( b_3, b_4, b_5 \) and \( b_6 \) are constants defined for each line \( k \). Below 0.7 hPa doppler broadening can be significant, so that in this case \( \gamma_k^{(v)} \) is modified to be

\[
\gamma_k^{(v)} = 0.535\gamma_k^{(v)} + \left( 0.217\left( \frac{\gamma_k^{(v)}}{\gamma_D} \right)^2 + \gamma_D \right)^{1.2} \quad \text{[GHz]},
\]

where \( \gamma_k^{(v)} \) is given by

\[
\gamma_k^{(v)} = 1 \times 10^{-3} b_3 (b_4 e(\epsilon_v(T_r), r)T_k^{66} + p_kT_k^{56}) \quad \text{[GHz]}
\]

and the Doppler width \( \gamma_D \) is

\[
\gamma_D = 1.46 \times 10^{-6} \nu_k^{(v)} T_r^{1/2} \quad \text{[GHz]}.
\]

The liquid water index of refraction is given by

\[
\bar{N}_W(T_r, \rho_w, \nu) = 1.5\rho_w \left\{ \frac{\epsilon_w - 1}{\epsilon_w + 2} \right\} \quad \text{[ppm]},
\]

which is a function of both the cloud water density \( \rho_w \) (g m\(^{-3}\)) and the complex permittivity of pure water \( \epsilon_w \). The complex permittivity of pure water, in turn, is approximated by a double-Debye fit to measured data,

\[
\epsilon_w = \epsilon_0 - f \left( \frac{\epsilon_0 - \epsilon_1}{\nu + i\gamma_1} + \frac{\epsilon_1 - \epsilon_2}{\nu + i\gamma_2} \right),
\]

with

\[
\begin{align*}
\epsilon_0 &= 77.66 + 103.3(T_r - 1), \\
\epsilon_1 &= 0.0671\epsilon_0, \\
\epsilon_2 &= 3.52 - 7.52(T_r - 1), \\
\gamma_1 &= 20.2 - 146.4(T_r - 1) + 316(T_r - 1)^2 \quad \text{[GHz]}, \\
\gamma_2 &= 39.8\gamma_1 \quad \text{[GHz]}.
\end{align*}
\]

Here, \( \epsilon_0 \) and \( \epsilon_1 \) are the static and high-frequency permittivities, respectively; \( \epsilon_2 \) is a model parameter which has a slight temperature dependence that has been removed in [Liebe et al., 1993] in order to avoid unphysical behavior for supercooled water above 100 GHz. As we do not consider liquid water effects below
freezing, we retain this dependence. The parameters $\gamma_1$ and $\gamma_2$ are two relaxation frequencies expressed in gigahertz; $\rho_w$ (g m$^{-3}$) is the liquid water density and $\nu$ is the radiation frequency in gigahertz.

The **index of refraction for ice** is given by

$$\tilde{N}(T_r, \rho_i, \nu) = 1.5 \left( \frac{\rho_i}{0.916} \right) \left( \frac{\varepsilon_i - 1}{\varepsilon_i + 2} \right) \text{ [ppm]},$$

where $\rho_i$ is the ice water density (g m$^{-3}$). The permittivity of ice, $\varepsilon_i$ is given as a function of electromagnetic frequency and reciprocal temperature by

$$\varepsilon_i = 3.15 + i \left( \frac{a_i}{\nu} + b_i \nu \right),$$

where

$$a_i = (T_r - 0.171) \exp [17.0 - 22.17T_r] \text{ [GHz]},$$

$$b_i = 1 \times 10^{-5} \left[ \left( \frac{0.233}{1 - 0.993/T_r} \right)^2 + \frac{6.33}{T_r} - 1.31 \right] \text{ [GHz$^{-1}$]}. $$

**5.2.3. Propagation Effects on the Stokes vector**

For a deterministic signal, the coefficients $a_h(r)$ and $a_v(r)$ in Eq (5.2.1 & 5.2.2) may be considered to be functions of distance, but for a non-deterministic signal these coefficients must be considered to be random variables, and in this case we are primarily interested in the second moments of the electric field vector. As such, we are concerned with the effect of the atmosphere on the Stokes vector,

$$\begin{bmatrix} I \\ Q \\ U \\ V \end{bmatrix} = \kappa \begin{bmatrix} \langle E_v E_v^* + E_h E_h^* \rangle \\ \langle E_v E_v^* - E_h E_h^* \rangle \\ 2\Re \langle E_v E_v^* \rangle \\ 2\Im \langle E_v E_v^* \rangle \end{bmatrix} = \begin{bmatrix} \langle a_v^2 + a_h^2 \rangle \\ \langle a_v^2 - a_h^2 \rangle \\ 2 \langle a_v a_h \cos(\delta_0) \rangle \\ 2 \langle a_v a_h \sin(\delta_0) \rangle \end{bmatrix},$$

where $\kappa = \frac{\lambda^2}{kB\eta_0 B}$ is a constant coefficient that depends upon the receiver bandwidth $B$, and the electromagnetic wavelength $\lambda$. The amplitude components are now random variables with a polarized and a non-polarized component. We now assume that the atmospheric attenuation rate $b(r)$ is not a random variable, so that the attenuation coefficients may be extracted from the ensemble averages. In this case, (5.2.3), (5.2.4), and (5.2.5) become, respectively,
\[
\begin{align*}
\frac{d\langle a_h^2 \rangle}{dr} &= -2b(r) \langle a_h^2(r) \rangle, \\
\frac{d\langle a_v^2 \rangle}{dr} &= -2b(r) \langle a_v^2(r) \rangle, \\
\frac{d\langle a_v a_h \rangle}{dr} &= -2b(r) \langle a_v a_h(r) \rangle.
\end{align*}
\] (5.2.6) (5.2.7) (5.2.8)

It follows directly from (5.2.6), (5.2.7), and (5.2.8) that the Stokes vector components satisfy the attenuation equations

\[
\frac{d}{dr} \begin{pmatrix} I(r) \\
Q(r) \\
U(r) \\
V(r) \end{pmatrix} = -2b(r) \begin{pmatrix} I(r) \\
Q(r) \\
U(r) \\
V(r) \end{pmatrix}
\]

The modified Stokes vector satisfies a similar equation:

\[
\frac{d}{dr} \begin{pmatrix} T_h(r) \\
T_v(r) \\
U(r) \\
V(r) \end{pmatrix} = -2b(r) \begin{pmatrix} T_h(r) \\
T_v(r) \\
U(r) \\
V(r) \end{pmatrix}
\]

5.2.4. Atmospheric Emission

The dispersion relation for electromagnetic radiation in free space is

\[\omega = ck = \frac{2\pi c}{\lambda},\]

where \(c\) is the speed of light in free space (m s\(^{-1}\)), \(\omega\) (rad Hz) and \(\lambda\) (m) are the electromagnetic radian frequency and wavelength, respectively. Thus,

\[
\frac{d\lambda}{d\omega} = \frac{2\pi c}{\omega^2} = \frac{(2\pi c)^2}{(2\pi c)^2} = \frac{\lambda^2}{2\pi c}.
\] (5.2.9)
To compute the atmospheric emission we assume that the atmosphere is characterized by an emissivity $\epsilon(r)$, and that its emission is unpolarized and well-approximated by the Rayleigh-Jeans approximation for blackbody emission multiplied by the (dimensionless) emissivity $\epsilon$

\[
I = \epsilon \frac{2\pi c k_b}{\lambda^4} T_p^4 \left( \frac{d\lambda}{d\omega} \right),
\]

where $I$ is the emitted power (specifically, the spectral radiance or brightness) with units (Wm$^{-2}$sr$^{-1}$Hz$^{-1}$), $T_p$ is the physical temperature. Using (5.2.9), (5.2.10) becomes

\[
I = \epsilon \frac{k_b}{\lambda^2} T_p
\]

In the above, $k_b$ is Boltzmann’s constant (1.38.10$^{-23}$ J K$^{-1}$). The emissivity accounts for the fact that the atmosphere emits much less radiation than a blackbody at the same physical temperature, and at L-band this emissivity is quite close to zero but not completely negligible compared with the signal variations owing to SSS variations. With this modified Rayleigh-Jeans form, the unpolarized atmospheric emission leads to the following ordinary differential equations for the Stokes and modified Stokes vector,

\[
\frac{d}{dr} \begin{pmatrix}
I(r) \\
Q(r) \\
U(r) \\
V(r)
\end{pmatrix} = \epsilon(r) \begin{pmatrix}
T_p(r) \\
0 \\
0 \\
0
\end{pmatrix},
\]

\[
\frac{d}{dr} \begin{pmatrix}
T_h(r) \\
T_v(r) \\
U(r) \\
V(r)
\end{pmatrix} = \epsilon(r) \begin{pmatrix}
T_p(r) \\
T_p(r) \\
0 \\
0
\end{pmatrix}.
\]

where now the emissivity $\epsilon = \epsilon(r)$ is a function of location along the propagation path, and the factor $k_b/\lambda^2$ does not appear because the Stokes elements implicitly contain this factor in their definition in terms of moments of the electric field.
5.2.5. Combining Attenuation and Emission

Combining the atmospheric attenuation and emission equations, we have

\[
\frac{d}{dr} \begin{pmatrix} I(r) \\ Q(r) \\ U(r) \\ V(r) \end{pmatrix} = -2b(r) \begin{pmatrix} I(r) \\ Q(r) \\ U(r) \\ V(r) \end{pmatrix} + \epsilon(r) \begin{pmatrix} 2T_p(r) \\ 0 \\ 0 \\ 0 \end{pmatrix},
\]

\[
\frac{d}{dr} \begin{pmatrix} T_h(r) \\ T_v(r) \\ U(r) \\ V(r) \end{pmatrix} = -2b(r) \begin{pmatrix} T_h(r) \\ T_v(r) \\ U(r) \\ V(r) \end{pmatrix} + \epsilon(r) \begin{pmatrix} T_p(r) \\ 0 \\ 0 \\ 0 \end{pmatrix}.
\]

Interestingly, with the assumptions we have made, the third and fourth Stokes parameters are affected by attenuation but not by emission.

Having established the ordinary differential equation the describes the change in the Stokes vector along the propagation path, we now turn to the problem of implementing a solution procedure that is practical for scene modeling. The straightforward approach is to ignore refraction and to integrate along linear paths through the atmosphere, and this is the approach we shall take.

5.2.6. Relating Attenuation to Emission

In general radiative transfer scattering may be important, so that attenuation may involve both absorption and scattering. At L-band, however, attenuation is dominated by absorption, even in the presence of cloud droplets and ice, so we may assume that the emissivity is equal to the power attenuation rate (twice the amplitude attenuation rate):

\[\epsilon(r) \approx 2b(r).\]

Given the similar forms for all Stokes vector components, in what follows we will discuss attenuation we will use the compact Stokes vector notation

\[\frac{d}{dr} \mathbf{T} = -2b(r)\mathbf{T} + b(r)\mathbf{T}_p,\]

where \(\mathbf{T}\) is the modified Stokes vector (with orthogonal linear components rather than I and Q) above) and
We will be concerned about propagation of radiation in the atmosphere at some zenith angle $\theta$ from the vertical direction, so we let $r = z \sec \theta$ be the path length in the propagation direction. We assume that, for a given location in the director cosine coordinates system of the SMOS antenna, we can compute the propagation path. As noted in the development above, in general this path will be curved if the propagation speed $c$ (related to the real part of $k$) is a function of position. However, we will neglect this curvature in what follows. At each point of the director cosine grid for which the line-of-sight intersects the earth, the zenith angle is assumed to be identical to the target incidence angle. At each point of the director cosine grid for which the line-of-sight does not intersect the Earth, the signal entering the radiometer will be equal to the sum of the atmospheric attenuated sky/sun/moon noise plus the integrated atmospheric emission along the path. In this case, the zenith angle can be computed directly in the Geographic Frame. In general, refraction should be taken into account when computing the integration path, but for L-band this refraction is expected to be negligible and so we have neglected it in our algorithm.

**5.2.7. Integration of the Radiative Transfer Equation**

As mentioned above, our main concern is attenuation and emission along the propagation path, where the amplitude attenuation coefficient is given by

$$b(r) = k N'' = 1 \times 10^{-6} \left( \frac{2 \pi \nu}{c} \right)^3 \exp \left\{ -\tilde{N} \right\} \text{ [Np m}^{-1}]$$

As before, $c$ is the speed of light in a vacuum in meters per second, $\nu$ is the electromagnetic frequency in hertz and $\tilde{N}$ is the sum of the individual indices of refraction for dry air, water vapor, cloud water, and cloud ice as given by in section 5.2.3:

$$\tilde{N}(T_r, e, p_d, \rho_w, \rho_i, \nu) = \tilde{N}_D(T_r, e, p_d, \nu) + \tilde{N}_W(T_r, e, p_d, \nu) + \tilde{N}_W(T_r, p_w, \nu) + \tilde{N}_I(T_r, \rho_i, \nu)$$

Here we have included explicit dependence of each index of refraction on atmospheric parameters $T_r, e, p_d, \rho_w, \rho_i$. In the forward model we typically take the atmospheric pressure $p$ (derived from geopotential height as a function of pressure), temperature $T$, relative humidity $r$, and cloud water mixing ratio fields $q_c$ from an atmospheric model such as the NCEP GFS. We use the NCEP GFS $1^\circ \times 1^\circ$ gridded fields of surface pressure (Pa), cloud water mixing ratio $q_w$ (kg/kg), temperature $T$ (K), relative humidity $r$ at geopotential height $z$. These analysis fields are available every 6 hours, and used to estimate the atmospheric attenuation and emission at 76 height levels from the surface to 15 km (200 m vertical grid spacing).

From the NCEP fields we compute the water vapor pressure $e$ from relative humidity $r$ and temperature $T$; we then compute the partial pressure of dry air, $p_d$ from $e$ and $p$. We assumed that water is entirely ice below freezing and entirely liquid above freezing, with liquid and ice water densities being $\rho_w = \rho_l = \rho_d q_c$, respectively, in their appropriate temperature ranges.

Having obtained $T, p_d, e, \rho_w$, and $\rho_i$, we then call four MPM93 functions implementing the above equations to compute the indices of refraction owing to dry air, water vapor, and liquid and ice water. The resulting refractive indices are then used to compute the attenuation rates in units of nepers per...
meter, which are then stored in another file for later spatial integration of the radiative transfer equation.

The solution procedure involves interpolation of the attenuation rates derived from the NCEP 6-hourly fields onto a uniformly spaced grid with Na=76 layers with layer bottom heights \( z_k \) ranging from \( z_s = 0 \) km through \( z_t = 15 \) km with a \( \Delta z = 200 \) m grid interval in geopotential height \( z \). The power attenuation rates, denoted by \( a_D(z), a_V(z), a_W(z) \) and \( a_I(z) \), are twice the corresponding amplitude attenuation rates given by \( b_D(z), b_V(z), b_W(z) \) and \( b_I(z) \). Using these power attenuation rates, we then compute the total absorption for each of these components along vertical paths from the bottom of the domain to each level. For simplicity, here we first form the total power attenuation rate, \( a(z) \), which is the sum of the individual attenuation rates of dry air, water vapor, and liquid and ice water:

\[
a(z) = 2b(r) = 2kN'' = a_D(z) + a_V(z) + a_W(z) + a_I(z)
\]

\[
= 2 \times 10^{-6} \left( \frac{2\pi\nu}{c} \right) 3 \left\{ \bar{N} \right\}
\]

\[
= 2 \times 10^{-6} \left( \frac{2\pi\nu}{c} \right) 3 \left\{ \bar{N}_D(T_r, e, p_d, \nu) + \bar{N}_V(T_r, e, p_d, \nu) + \bar{N}_W(T_r, \rho_w, \nu) + \bar{N}_I(T_r, \rho_i, \nu) \right\}
\]

The integrated attenuation from the surface at height \( z_s \) to height \( z \) is

\[
A^{(u)}(z) = A(z_s, z) = \int_{z'=z_s}^{z} a(z') dz',
\]

where the superscript denotes integration from below up to height \( z \). Similarly, the integrated attenuation from height \( z \) to the top of the atmosphere is

\[
A^{(d)}(z) = A(z, z_t) = \int_{z'=z}^{z_t} a(z') dz' \quad [\text{Np}],
\]

where the superscript denotes integration from above down to height \( z \). The corresponding power transmittances for radiation propagating upward from below and downward from above are, respectively

\[
\tau^{(u)}(z) = \tau(z_s, z) = \exp \left[ -A^{(u)}(z) \right] = \exp \left[ \int_{z'=z_s}^{z} a(z') dz' \right].
\]

\[
\tau^{(d)}(z) = \tau(z, z_t) = \exp \left[ -A^{(d)}(z) \right] = \exp \left[ \int_{z'=z}^{z_t} a(z') dz' \right].
\]

In discrete form the integrated upward and downward attenuations in nepers are, respectively,

\[
A^{(u)}(z_m) = \sum_{k=1}^{m} a(z_k) \Delta z \quad [\text{Np}],
\]

\[
A^{(d)}(z_m) = \sum_{k=m}^{N_a} a(z_k) \Delta z \quad [\text{Np}],
\]
where \( A^{(u)}(z_m) \) and \( A^{(d)}(z_m) \) have units of nepers since they are attenuation rates integrated over small height ranges. Here the subscript \( m \) is an index into the vector of vertical layers that form a discrete vertical grid with layer center heights \( z_m \). Except at near grazing zenith angles, an arbitrary path will not traverse a horizontal distance of more than a few kilometers, and if we assume that the atmosphere on this spatial scale is horizontally uniform, then the total attenuation along a path at zenith angle \( \theta \) may be obtained from the preceding total attenuation values by multiplication by \( \sec \theta \). The resulting zenith transmissivity between the surface and layer \( k \) is then

\[
\tau^{(u)}(z_m) = \tau(z_s, z_m) = \exp \left[ -A^{(u)}_m \right].
\]

In terms of the individual contributions by dry air, water vapor, cloud water and cloud ice, the total transmissivity from the surface to the top of layer \( z_m \) is

\[
\tau^{(u)}(z_m) = \tau(z_s, z_m) = \exp \left[ -A^{(u)}(z_m) \right] = \tau_D^{(u)}(z_m)\tau_V^{(u)}(z_m)\tau_W^{(u)}(z_m)\tau_I^{(u)}(z_m),
\]

while the total transmissivity from the top of the atmosphere to the base of layer \( z_m \) is

\[
\tau^{(d)}(z_m) = \tau(z_s, z_m) = \exp \left[ -A^{(d)}(z_m) \right] = \tau_D^{(d)}(z_m)\tau_V^{(d)}(z_m)\tau_W^{(d)}(z_m)\tau_I^{(d)}(z_m).
\]

The transmissivities at zenith angle \( \theta \) are obtained by raising these zenith transmissivities to the negative \( \sec \theta \) power:

\[
\tau^{(u)}(z_m, \theta) = \tau(z_s, z_m, \theta) = \exp \left[ -A^{(u)}(z_m) \sec \theta \right] = \left[ \tau_D^{(u)}(z_m)\tau_V^{(u)}(z_m)\tau_W^{(u)}(z_m)\tau_I^{(u)}(z_m) \right]^{\sec \theta},
\]

\[
\tau^{(d)}(z_m, \theta) = \tau(z_s, z_m, \theta) = \exp \left[ -A^{(d)}(z_m) \sec \theta \right] = \left[ \tau_D^{(d)}(z_m)\tau_V^{(d)}(z_m)\tau_W^{(d)}(z_m)\tau_I^{(d)}(z_m) \right]^{\sec \theta}.
\]

To complete the model we must account for atmospheric emission. The (assumed to be unpolarized) brightness temperature emitted by the atmosphere per unit height interval at some height \( z \) above the ground (i.e., a brightness temperature rate) is, assuming that the atmosphere is in thermal equilibrium,

\[
\dot{T}_{ba}(z) = a(z)T_p(z),
\]

where \( T_p(z) \) is the physical temperature at height \( z_k \) and \( \dot{T}_{ba}(z_k) \) is a brightness temperature rate with units of kelvin per meter. The total brightness temperature rate of atmospheric emission at the height \( z_k \) is, considering all components,

\[
\dot{T}_{ba}(z_k) = \dot{T}_{bd}(z_k) + \dot{T}_{bw}(z_k) + \dot{T}_{bw}(z_k) + \dot{T}_{bi}(z_k).
\]

We can then vertically integrate this total brightness temperature rate to derive the upwelling attenuated atmospheric brightness temperature at the top of the atmosphere:
which in discrete form becomes

$$T_{ba}^{d\alpha} = \sum_{k=1}^{m} \tau^{(d)}(z_k)T_{bak} = \sum_{k=1}^{m} \tau(z_k, t_i)T_{bak} = \sum_{k=1}^{m} \tau(z_k, t_i)a(z_k)T_p(z_k)\Delta z,$$

where

$$T_{bak} \approx a(z_k)T_p(z_k)\Delta z$$

is the unattenuated incremental brightness temperature of emission in layer $k$ (with units of kelvin). Similarly, the zenith downwelling brightness temperature at the bottom of the atmosphere is

$$T_{ba}^{ boa} = \int_{z'=z_s}^{z_t} \hat{T}_{ba}(z') \exp \left[ - \int_{z''=z_s}^{z'} a(z'') d z'' \right] d z'$$

which in discrete form becomes

$$T_{ba}^{ boa} = \sum_{k=1}^{m} \tau^{(u)}(z_k)T_{bak} = \sum_{k=1}^{m} \tau(z_s, z_k)T_{bak} = \sum_{k=1}^{m} \tau(z_s, z_k)a(z_k)T_p(z_k)\Delta z.$$

Although $T_{ba}^{ boa}$ and $T_{ba}^{ boa}$ do not differ much, they are not identical.

References


5.3. Corrections for Ocean surface scattered sky noise

Modeling studies conducted by several teams prior to SMOS launch indicated that the downwelling celestial radiations at L-band that are scattered back by the ocean surface toward the upper hemisphere can be a source of brightness contamination affecting the quality of sea surface salinity retrieval (e.g. see Tenerelli et al., 2008; Reul et al., 2008).
For sun-synchronous polar-orbiting satellite measurements of upwelling L-band radiation over the ocean, like with SMOS, this so-called sky noise depends strongly on pass direction (ascending or descending), time of year and surface roughness (wind speed).

![Image of scattered celestial noise](https://example.com/sky_noise_images)

**Figure:** Maximum unpolarized scattered celestial noise \( \frac{1}{2}(Tv + Th) \) over all measurements of each dwell line for (a) descending and (b) ascending swaths. The wind speed is 7 m.s\(^{-1}\), and the downwind direction is 0\(^\circ\). The Kudryavtsev wave spectrum and the KA scattering model are used to compute the scattered signal. Solutions are expressed in kelvin.

Based upon the modeling studies for SMOS sensor and as illustrated in the above Figure, the impact is expected to be strongest for descending passes from August to October because for these passes the reflections of the instrument viewing directions over the field of view tend to lie along the galactic equator where L-band galactic emission is maximum.

The forward model we used to correct for Ocean surface scattered galactic radiation is the same than the one used in the operational ESA Level 2 processor. It is based on a Kirchoff approximation for the rough sea surface scattering model. The algorithm is described in detail in the operational ESA Level 2 processor ATBD.

SSS retrievals produced by DPGS using that model show along-track strips of low SSS in descending orbits that are aligned with the expected maximum in galactic noise contribution to the brightness temperatures. If the contribution to the brightness temperatures measured by SMOS from celestial sky noise is underpredicted, the residual brightness temperatures will be overpredicted and the resulting sea surface salinity will be underpredicted (because brightness temperature increases as surface salinity decreases).

Examination of the SMOS brightness temperatures and comparison with the scattering model reveals deficiencies in the modeling of the celestial sky noise that are at the origin of these strips of spuriously low SSS: in general we underestimate the scattered energy from the galactic equator, meaning that our sea surface scattering model is overpredicting the roughness effect. An adapted correction is understudy to improve that correction. To minimize these effects for the present Level 3 products, we used a forward scattering model solution at a fixed wind of 3 m/s.
5.4. Evaluation of First Stokes parameter at the surface

5.4.1. First Stokes parameter at the antenna level

An additional source of earth surface emitted brightness modification at L-band as measured from space is the polarization mixing (Faraday rotation), due to the electromagnetic wave propagation through the ionosphere in the presence of the geomagnetic field (Skou, 2003). It can be either modeled from the knowledge of the ionospheric Total Electron Content (TEC) and magnetic field obtained from an external source or avoided by using the first Stokes parameter $I = T_H + T_V$, which is basically invariant by rotation.

For our sea surface salinity retrieval algorithm, we chose here this alternative option and estimated the first Stokes brightness temperature parameter from the reconstructed XX and YY antenna polarization brightness temperature fields expressed in the antenna director cosine coordinates $(\xi, \eta)$:

$$I_{\text{TOA}}(\xi, \eta) = T_{\text{xx}}(\xi, \eta) + T_{\text{yy}}(\xi, \eta) = T_H(\xi, \eta) + T_V(\xi, \eta).$$

SMOS instrument is not measuring both linear polarization (xx) and (yy) at the same acquisition time. In dual polarization mode the Level 1B brightness temperature frequencies are delivered successively in XX and YY polarisations (namely $T_{\text{xx}}$ and $T_{\text{yy}}$) every $\Delta t=1.2$ ms, the antenna integration time. In full polarisation mode, visibilities are acquired following a complex temporal cycle of rotating polarizations. To build up a First Stokes parameter, we therefore combined successively acquired linearly polarized data following a "scene" concept. In that concept, we basically assume that successively acquired linear polarization correspond to the same brightness temperature scene (neglecting the antenna boresight displacement over $\Delta t$).

In dual polarization mode, a scene is defined by the combination of two successive snapshots separated by the antenna integration time $\Delta t$, so that the the first stokes parameter at the antenna level for the scene indexed number "i" :$\text{scene}_i$ corresponding to a "mean" acquisition time $t_i$ will be obtained following:

$$I_{\text{TOA}}(\xi, \eta, \text{scene}_i) = T_{\text{xx}}(\xi, \eta, t_i) + T_{\text{yy}}(\xi, \eta, t_i + \Delta t)$$

In full polarization mode, the Level 1B brightness temperature frequencies are delivered in the following order:

Receivers in all arms in X-pol for 1.2 s:

1) XX-long

References:


Receivers in X-pol in 2 arms and in Y-pol in the third arm rotating for 1.2 s:
2) XX-short
3) XY-real
4) XY-imaginary
Receivers in all arms in Y-pol for 1.2 s:
5) YY-long
Receivers in Y-pol in 2 arms and in X-pol in the third arm rotating for 1.2 s:
6) YY-short
7) XY-real
8) XY-imaginary

To deal with the complexity of the SMOS polarimetric information in full polarization mode, we introduce the ‘scene’ here consists of a collection of 4 snapshots yielding the complete Stokes vector (XX,YY,XYR,XYI).

The individual snapshots corresponding to a given complete Stokes vector scene may be obtained from the following index mappings, which provide the set of snapshot indices for each scene index. These scene index mappings are stored in the snapshots substructure within each analysis file and have the following names:

Our algorithm reads the Level 1B DBL file and loops through all snapshots in the order in which they appear in the file. The snapshots are assumed to be in order of increasing time.

The code reads successive snapshots until a complete Stokes vector is obtained, without regard to whether or not particular snapshots correspond to mixed or pure polarization modes of the instrument. As each snapshot is read, any snapshot already read that is being considered to be part of the next complete Stokes vector is discarded if the time difference between this snapshot and the one just read exceeds 5 seconds.

In this manner complete Stokes vector scenes consisting of successive snapshots are created. Note that, given the typical ordering of snapshots in the data files, each scene will typically consist of either a mixed-XX and full-YY snapshot or a full-XX and mixed-XX snapshot.

5.4.2. Collection of Auxiliary Parameters

For a given scene $i$, a series of geometrical and auxiliary geophysical parameter are determined at each antenna cosine director coordinate $\xi, \eta$.

The geometrical parameters are obtained using Earth CFI software and include:

- Latitude of the earth target: $\text{lat}(\xi, \eta, \text{scene}_i)$
- Longitude of the earth target: $\text{lon}(\xi, \eta, \text{scene}_i)$
- Earth incidence angle at the target: $\theta(\xi, \eta, \text{scene}_i)$
- Look direction azimuth at the target $\phi(\xi, \eta, \text{scene}_i)$
- Right ascension of the earth specularly reflected path in the B1950 sky coordinate system $\text{ra}(\xi, \eta, \text{scene}_i)$
- Declination of the earth specularly reflected path in the B1950 sky coordinate system $\text{dec}(\xi, \eta, \text{scene}_i)$
• cosine director coordinate of the sun $\xi_{\text{sun}}, \eta_{\text{sun}} (\text{scene}_i)$
• cosine director coordinate of the moon $\xi_{\text{moon}}, \eta_{\text{moon}} (\text{scene}_i)$
• cosine director coordinate of the earth specularly reflected path towards the sun $\xi_{\text{spec}}^{\text{sun}}, \eta_{\text{spec}}^{\text{sun}} (\text{scene}_i)$
• cosine director coordinate of the earth specularly reflected path towards the moon $\xi_{\text{spec}}^{\text{moon}}, \eta_{\text{spec}}^{\text{moon}} (\text{scene}_i)$

• a landmask flag obtained from USGS mask
• a sea-ice flag from SSM/I

To detect pixels potentially affected by sea ice, we used the PSI-SSMI level 2 product delivered at Ifremer/cersat which contains, for both north and south poles, polar stereographic 12.5 km resolution grids of sea ice concentration from the 85 GHz channel of SSM/I on DMSP. The daily maps are processed from the daily brightness temperature maps from NSIDC. The Artist Sea Ice (ASI) algorithm developped at University of Bremen (Germany) is used to processed daily sea ice concentration maps at 12.5 km resolution. A sea ice flag is raised as soon as sea ice concentrations values higher than 10% are detected.

Given the "mean" aquisition time $t_i$ of the scene $\#i$ and the geolocated director cosine coordinates $[\text{lat}(\xi, \eta), \text{lon}(\xi, \eta)]$ for that scene, the geophysical auxiliary parameters are spatio-temporally interpolated either from the SMOS ECMWF auxilliary products and/or from the NCEP GFS 6-hourly products. They include:

• sea level pressure: $P(\xi, \eta, \text{scene}_i)$,
• cloud water mixing ratio at geopential z: $q_w(\xi, \eta, \text{scene}_i, z)$,
• atmospheric temperature at geopential z: $T(\xi, \eta, \text{scene}_i, z)$
• relative humidity at geopential height: $r(\xi, \eta, \text{scene}_i, z)$
• zonal wind speed component at 10 meter height: $u_{10}(\xi, \eta, \text{scene}_i)$
• meridional wind speed component at 10 meter height: $v_{10}(\xi, \eta, \text{scene}_i)$
• surface wind speed at 10 meter height: $w_{10}(\xi, \eta, \text{scene}_i)$
• wind direction azimut: $\varphi_{w}(\xi, \eta, \text{scene}_i)$
• wind friction velocity: $U_{c}(\xi, \eta, \text{scene}_i)$
• wave age: $\Omega(\xi, \eta, \text{scene}_i)$
• sea surface temperature: $T_s(\xi, \eta, \text{scene}_i)$
• Air temperature at 2m height: $T_{\text{air}2m}(\xi, \eta, \text{scene}_i)$
• WOA2005 sea surface salinity climatology: $SSS_{\text{clim}}(\xi, \eta, \text{scene}_i)$
• ECMWF WAM estimate for the Charnock parameter $\text{chnk}(\xi, \eta, \text{scene}_i)$
• ECMWF WAM model peak period of the 1D spectrum: $T_p(\xi, \eta, \text{scene}_i)$
• ECMWF WAM model drag coefficient for wind waves: $C_D(\xi, \eta, \text{scene}_i)$
• ECMWF WAM model peak period for wind waves: $T_{\text{pow}}(\xi, \eta, \text{scene}_i)$
• ECMWF WAM model mean square slope: $\text{mss}(\xi, \eta, \text{scene}_i)$
• ECMWF WAM model significant wave height: $H_s(\xi, \eta, \text{scene}_i)$

We use a linear interpolation in time and bi-linear interpolation in space to co-register these auxilliary geophysical data with SMOS data.

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5.4.3. Total power emissivity at the surface level

The reconstructed First Stokes brightness temperatures estimated for each individual scene are first corrected to account for the image reconstruction-induced spatial biases using the Ocean Target Transformation (OTT) generated as described in §5.1:

\[ \bar{I}_{TOA}^{5.4.3}(\xi, \eta, \text{scene}) = I_{TOA}^{5.4.3}(\xi, \eta, \text{scene}) - [\text{OTT}_s(\xi, \eta) + \text{OTT}_t(\xi, \eta)] \]

Given the auxiliary atmospheric and geometric parameters at each node of the antenna cosine director coordinate system, we evaluate for each scene the atmospheric contributions using the forward model as described in §5.2:

- \( \tau^u(z_r, \xi, \eta) \): the total atmospheric transmissivity from the surface to the top of atmosphere (given at height \( z_r = 15 \) km),
- \( T_{ba}^{5.4.3}(\xi, \eta) \): the upwelling attenuated atmospheric brightness temperature at the top of the atmosphere,
- \( T_{bo}^{5.4.3}(\xi, \eta) \): the downwelling atmospheric brightness temperature at the bottom of the atmosphere,

The First Stokes brightness temperatures estimates at the antenna level \( \bar{I}_{TOA}^{5.4.3} \) can be expressed as:

\[ \bar{I}_{TOA}^{5.4.3}(\xi, \eta, \text{scene}) = T_{bo}^{5.4.3}(\xi, \eta) + \tau^u(\xi, \eta) \left[ I_{scat}(\xi, \eta) + I_{surf}(\xi, \eta) + |R(\xi, \eta)|^2 T_{ba}^{5.4.3}(\xi, \eta) \right] \] (5.4.3.1)

where \( I_{scat}(\xi, \eta) \) is the scattered sky noise signal estimated from the galactic correction model as described in §5.3, \( I_{surf}(\xi, \eta) = I_{\text{flat}}(\xi, \eta) + \Delta I_{\text{roug}}(\xi, \eta) \) is the total power surface brightness temperature including the contribution from the perfectly smooth sea emission \( (I_{\text{flat}}(\xi, \eta)) \) and from the rough sea surface emission \( (\Delta I_{\text{roug}}(\xi, \eta)) \). The downwelling atmospheric brightness temperature at the bottom of the atmosphere \( T_{ba}^{5.4.3} \) is reflected by the ocean surface towards the sensor. In theory, to evaluate the latter atmospheric contribution, one shall take into account the scattering effect of these downwelling radiations by the rough sea surface. Here we however neglect this effect which is very small and only consider the specular reflection of the downwelling atmospheric brightness temperature at the ocean surface. The latter is evaluated using the Fresnel reflectivity at incidence angle \( \theta(\xi, \eta) \) using the Klein and Swift (1977) dielectric constant model assuming that the sss at lat(\( \xi, \eta \)); lon(\( \xi, \eta \)) is given by the WOA2005 climatology and using the sst from ECMWF.

The total power surface emissivity for each node \((\xi, \eta)\) of the director cosine coordinate system and for each individual scene is then obtained from (5.4.3.1) using:

\[ e_{surf}(\xi, \eta, \text{scene}) = \left[ \frac{\bar{I}_{TOA}^{5.4.3}(\xi, \eta) - T_{ba}^{5.4.3}(\xi, \eta)}{\tau^u(\xi, \eta)} - I_{scat}(\xi, \eta) - |R(\xi, \eta)|^2 T_{ba}^{5.4.3}(\xi, \eta) \right] / \text{sst}(\xi, \eta) \] (5.4.3.2)

5.5. Incidence angle binning and gridded First Stokes multi-angular surface Emissivities

For each pass-type separately (ascending or descending), smos data acquired over one day are collected together. The surface emissivities \( e_{surf}(\xi, \eta, \text{scene}) \) for all scenes in that ensemble are...
estimated using (5.4.3.2) and considering only data with director cosine coordinates within the Alias Free Field of View domain.

The data are further classified as function of their incidence angle values $\theta(\xi, \eta)$. Classification is performed over the ten following incidence angle bins ranging from $12.5^\circ$ to $62.5^\circ$ with $\pm 2.5^\circ$ width:

$$
\begin{align*}
\theta_{\text{bin}}^1 &= [12.5^\circ, 17.5^\circ] \text{ with central incidence angle } \theta_{\text{bin}}^1 = 15^\circ \\
\theta_{\text{bin}}^2 &= [17.5^\circ, 22.5^\circ] \text{ with central incidence angle } \theta_{\text{bin}}^2 = 20^\circ \\
\theta_{\text{bin}}^3 &= [22.5^\circ, 27.5^\circ] \text{ with central incidence angle } \theta_{\text{bin}}^3 = 25^\circ \\
\theta_{\text{bin}}^4 &= [27.5^\circ, 32.5^\circ] \text{ with central incidence angle } \theta_{\text{bin}}^4 = 30^\circ \\
\theta_{\text{bin}}^5 &= [32.5^\circ, 37.5^\circ] \text{ with central incidence angle } \theta_{\text{bin}}^5 = 35^\circ \\
\theta_{\text{bin}}^6 &= [37.5^\circ, 42.5^\circ] \text{ with central incidence angle } \theta_{\text{bin}}^6 = 40^\circ \\
\theta_{\text{bin}}^7 &= [42.5^\circ, 47.5^\circ] \text{ with central incidence angle } \theta_{\text{bin}}^7 = 45^\circ \\
\theta_{\text{bin}}^8 &= [47.5^\circ, 52.5^\circ] \text{ with central incidence angle } \theta_{\text{bin}}^8 = 50^\circ \\
\theta_{\text{bin}}^9 &= [52.5^\circ, 57.5^\circ] \text{ with central incidence angle } \theta_{\text{bin}}^9 = 55^\circ \\
\theta_{\text{bin}}^{10} &= [57.5^\circ, 62.5^\circ] \text{ with central incidence angle } \theta_{\text{bin}}^{10} = 60^\circ
\end{align*}
$$

The $e_{\text{surf}}(\text{lat}(\xi),\text{lon}(\eta),\text{scene})$ data collected over each day are vectorized and classified in each incidence angle bin $\theta_{\text{bin}}^i$. They are further spatially averaged separately for each pass-type over a $0.25^\circ \times 0.25^\circ$ global grid.

Two examples of daily gridded surface emissivity fields at $\theta_{\text{bin}}^5 = [32.5^\circ, 37.5^\circ]$ are shown in the figure here below:

**Figure**: Examples of daily gridded surface emissivities $e_{\text{surf}}(\text{lat}, \text{lon}, \theta_{\text{bin}}^i, \text{day of year})$ at the incidence angle bin $\theta_{\text{bin}}^5 = [32.5^\circ, 37.5^\circ]$ with central incidence angle $\theta_{\text{bin}}^{5c} = 35^\circ$ and for day of year 2010 #191. Left: ascending passes; right: descending passes.

From this processing step, we obtained time series of estimated surface emissivities at global scale in each ten incidence angle bins and for each pass-type:

$$
\begin{align*}
e_{\text{surf}}^{\text{asc}}(\text{lat}, \text{lon}, \theta_{\text{bin}}^i, \text{day}) & \quad \text{for ascending passes} \\
e_{\text{surf}}^{\text{desc}}(\text{lat}, \text{lon}, \theta_{\text{bin}}^i, \text{day}) & \quad \text{for descending passes}
\end{align*}
$$
where "doy" stands for "day of year".

Co-registered key geophysical information characterizing the surface state were as well collected in the same manner separately for each pass (to account for 6pm versus 6am geophysical changes in the same day, e.g. wind speed diurnal changes) such as:

- sea surface temperature $T_{s}^{Asc, Desc}(lat, lon, doy)$,
- 10 meter height surface wind speed $W_{10}^{Asc, Desc}(lat, lon, doy)$,
- significant wave height $H_{s}^{Asc, Desc}(lat, lon, doy)$,
- and other parameters listed in §5.4.2.
6. Spatio-temporal Variability Analysis

6.1 Sources of Spatio-temporal variability

Spurious data can be found in the temporal time series of $e_{surf}$ at a given earth grid node (lat,lon) for some half orbits within the year during which instrument calibration sequences were performed (e.g. electrical stability test, short term calibration) or in cases for which very strong radio frequency interferences contaminated the observations. As a first processing step, we therefore filtered out from the emissivity data time serie fields, unphysical ocean emissivities for both passes. The latter are defined by:

\[
\text{Filter 1: } e_{surf} > 1 \text{ or } e_{surf} \leq 0
\]

After applying this first-step filter the annual mean and standard deviation of the $e_{surf}$ data were evaluated. The resulting surface emissivity variability maps for each incidence angle bin an each pass are illustrated in the figure below.
As illustrated in the above Figure, most locations over the global ocean are associated with an L-band surface emissivity that exhibits an annual standard deviation (asd) less or equal than ≈0.03. The signal with the world largest sea surface salinity seasonnal variability is the amazon plume area: it is associated with a maximum annual std of the surface L-band emissivity on the order of 0.04 at the river mouth. Roaring Forties and Furious Fifties area with highly variable roughness states exhibits asd of the surface emissivity lower than 0.03.

Nevertheless, large oceanic zones within SMOS data exhibits surface emissivity asd that exceeds 0.04. The spatial patterns of these highly variable zones is strongly dependent on the observing incidence angle value and on the pass-type (asc versus desc). Most of these highly variable zones are indeed associated with RFI contamination. In heavily RFI contaminated zones, such as the arabian sea, annual std as high as 0.2 can be encountered.

Therefore, several sources can explain the observed spatio-temporal variability in the SMOS estimates of surface emissivity $e_{surf}$. The major sources of variability in these signals include:

- Radio Frequency Interferences impact on the reconstructed brightness temperatures,
- Ascending/Descending biases (such as the one induced by errors in the instrument thermal response monitoring, solar & galactic radiations,...),
- Land & ice Contamination,
- Solar radiation impacts,
- errors in the correction of atmospheric and sea surface scattered sky radiation contributions, and,
- Sea surface Geophysical variability (sea surface salinity, temperature & roughness)

In order to perform a robust Sea surface Salinity inversion from $e_{surf}$ estimates, it is mandatory to best isolate the geophysical sources of surface variability in $e_{surf}$ data which include dielectric variability (induced by sst and sss changes) but also sea surface roughness induced variability. The other perturbing sources of variability (RFI, asc/desc biases, land & ice effects, solar and sky
radiation impacts, ...) shall be detected and removed/filtered from the data set. We filtered out these spurious signals through an analysis of the statistics of the $e_{surf}$ data over the complete year. The filtering methodology is based upon the specific signatures of non-geophysical signal and rely on some data internal consistency check tests.

In this section, we first review the main sources of variability of $e_{surf}$ and their characteristics as identified when stacking together one year of acquired data. In a second step, we present the algorithm which we used to identify surface emissivity samples at a given location on earth that deviate anomalously from the average of their neighbors and/or of an expected geophysical variability as function of space, time, earth incidence angle and pass types (ascending/descending).

### 6.2 The Radio Frequency Interference Issue

One of the most important unwanted source of variability in the $e_{surf}$ signals is associated with contaminations from radio frequency interferences (RFI). SMOS multi-angular brightness temperature measurements at 1.4 GHz are strongly affected by RFI from radar networks, TV and radio links in what shall be a protected band. These interferences are numerous over land in Europe and Asia, but can be also encountered in some other areas of Africa, America and Greenland and in some numerous islands over the world. Over the oceans, the signals emanating from land sources can extend very far away from the coasts and have dramatic consequences on the accuracy of sea surface salinity remote sensing in some key oceanic areas like the north atlantic, north pacific, Asia Coastline and north indian oceans. The signature of RFI in SMOS data is highly variable in time and space and strongly depends on the instrument probing polarization and observation angles. Because of the interferometric principle, local strong RFI sources in the physical space can contaminate a very extended area in the Fourier domain of the synthetic antenna and contaminate large portion of the SMOS reconstructed brightness temperature images (a point source => a star in the Fourier domain). In particular, RFI sources located in the aliased regions of the image can impact the data in the (extended) alias-free field of view. Ideally, detection and Mitigation techniques of these spurious signals in SMOS data shall therefore be performed from the raw data at the visibility level prior image reconstruction and shall consider instantaneous acquisitions (snap-shot information) and deal with the whole field of view images. Nevertheless, because of the strong amplitude of the RFI contamination with respect geophysical signal over the ocean, simple detection algorithm can be applied to the reconstructed brightness temperature data, identifying samples that deviate anomalously from the average of their neighbors in space, time and probing angles. Collecting acquired data over the full year 2010, we were in a position to re-analyze the spatio-temporal characteristics of these signals and their varying signature as function of the instrument probing configuration (incidence angle, ascending versus descending passes). A global RFI analysis over the world ocean was performed from that data ensemble. The large number of data acquired during the one year time series at a given location on earth allowed us to clearly establish robust threshold detection criteria for these contaminations.

As illustrated in Figure (?), most locations were the surface emissivity annual standard deviation exceed 0.04 is associated with RFI contaminated zones. The geophysical signal with largest variability is the amazon plume with a maximum annual std on the order of 0.04 as well. In heavily RFI contaminated zones, such as the arabian sea, annual std as high as 0.2 can be encountered.

A particularity of RFI signatures in SMOS data is their strong dependence on incidence angle value and track type (ascending/descending).
6.1.1. Example of some RFI in the Indian Ocean: Isolated Outlier Detection

Here, we show an example of the surface emissivity time series along the year 2010 at a given location in the South West Indian Ocean and for the incidence angle bin $\theta_{2c}^{bin} = 20^\circ$.

Figure: time series of the surface emissivity at a given location in the Indian Ocean and for the incidence angle bin $\theta_{2c}^{bin} = 20^\circ$ for ascending orbits (blue) and descending orbits (red). The dashed lines are indicated the mean annual values and the mean±3 times the annual standard deviation.

In that example, we see that the surface emissivities for ascending and descending orbits at $20^\circ$ incidence angle exhibit similar annual average values ($\mu_{e_{surf}}^A = \mu_{e_{surf}}^D \approx 0.62$) at the chosen point $\tilde{x}$ but clearly show differing annual standard deviation. The latter is significantly higher for the descending orbits (red color dots and lines) than for ascending ones (blue ones): $\sigma_{e_{surf}}^D(x, \theta_i = \theta_{2c}^{bin}) > \sigma_{e_{surf}}^A(x, \theta_i = \theta_{2c}^{bin})$. As illustrated, this is due to an isolated outlier value in descending orbit around day of year ~305, with surface emissivity $e_{surf}^D < \mu_{e_{surf}}^D(x, \theta_i = \theta_{2c}^{bin}) - 3 \sigma_{e_{surf}}^D(x, \theta_i = \theta_{2c}^{bin})$.

Therefore, isolated outliers most likely generated by local in space and time RFI events can be first detected from annual statistics determined for each incidence angle bin and for each orbit type, separately.

6.1.2. Example of some South Atlantic RFI Signals East of Argentina

In most cases however, RFI events are not fully local in time and space and can affect the salinity retrieval over an extended spatial domain and over a long period of time. However, RFI-induced signatures show a strong variability as function of the multi-angular viewing geometry of SMOS but also depend, at a given incidence angle, on the varying viewing geometry from ascending orbits to descending orbits. As illustrated in the below examples in the South Atlantic Ocean, these properties can be used to detect anomalous behaviors in the surface emissivity time series and to remove them from the dataset prior the surface emissivity analysis to retrieve surface salinity.
6.1.2.1. Mono-angular detection and filtering

As a next processing step for RFI detection and filtering, we consider the detection of spurious RFI-induced signals based solely on the different behaviours in between ascending and descending orbits data at the same incidence angle and for the same ocean location.

![Example blow-up map east of Argentina](image)

Figure: Examples blow-up map east of Argentina of the annual standard deviation of surface emissivities at an incidence angle of $\theta_{\text{inc}} = 20^\circ$ (upper plots) and of the annual mean emissivity (bottom plots). Right plots are for descending orbits and left plots are for ascending ones. The black line is the location of a latitudinal section (S1) through the RFI contaminated zone and the black circle (denoted P1) illustrates a chosen location within that contaminated area.

In the example shown in the above figure, we have a significant difference in the annual standard deviation and annual mean values of the surface emissivity at $\theta_{\text{inc}} = 20^\circ$ in between ascending and descending orbits over an extended area east of Argentina, north west of the South Georgia island.

A latitudinal section of the emissivity field statistics crossing that contaminated area is shown in the next figure left plot. As illustrated, from $\approx 52^\circ$S to $47^\circ$S, the mean and annual standard deviation of the surface emissivity in ascending orbits strongly depart from their descending orbits counterparts. The time series at point 1, located in that contaminated area is shown in the Figure right plot.
Figure: Left: Annual average surface emissivity along the latitudinal Section #1 ±1 std. Thick blue (Ascending orbits); Thick red (Descending orbits). Right: temporal time series at Point P1. Thick dashed horizontal lines are showing the annual mean value and the thin ones are showing the mean ±3 std.

In that case, it can be assumed that the surface emissivity data set from the orbit-type that exhibit the lower annual standard deviation value in between ascending and descending orbits are measured for passes that are significantly less perturbed by RFI contaminations (in our particular example: the descending ones since $\sigma_{e_{surf}}^d(\tilde{x}, \theta_i = \theta_{bin}^{2c}) > \sigma_{e_{surf}}^a(\tilde{x}, \theta_i = \theta_{bin}^{2c})$). A second outlier detection and filtering step can therefore be defined in that case by selecting only those contaminated ascending orbit emissivity data $e_{surf}^a(x, t, \theta_i)$ acquired during the year that fall beyond the range of observed descending orbits data, such that:

$$\mu_{e_{surf}}^d(\tilde{x}, \theta_{bin}^{2c}) - 3 \sigma_{e_{surf}}^d(x, \theta_{bin}^{2c}) \leq e_{surf}^a(\tilde{x}, t, \theta_{bin}^{2c}) \leq \mu_{e_{surf}}^d(\tilde{x}, \theta_{bin}^{2c}) + 3 \sigma_{e_{surf}}^d(x, \theta_{bin}^{2c})$$

where $\mu_{e_{surf}}^d(\tilde{x}, \theta_{bin}^{2c})$ and $\sigma_{e_{surf}}^d(x, \theta_{bin}^{2c})$ are the annual mean and standard deviation of emissivities in descending orbits at incidence angle $\theta_{bin}^{2c}$, respectively and $\mu_{e_{surf}}^a(\tilde{x}, \theta_{bin}^{2c})$ and $\sigma_{e_{surf}}^a(x, \theta_{bin}^{2c})$ are the annual mean and standard deviation of emissivities in ascending orbits at incidence angle $\theta_{bin}^{2c}$, respectively.

The result of applying that filter is shown in the next figure here below:
Figure: Left: Annual average surface emissivity along the latitudinal Section #1 ±1 std after filtering ascending orbit data following a constraint based on the descending orbit data annual statistics. Thick blue (Ascending orbit); Thick red (Descending orbit). Right: temporal time series at Point P1 after filtering ascending passes data. The thick lines indicate the annual mean and the thin lines indicate the new annual mean±3 the standard deviation.

As shown, the previous filtering operation significantly reduced the biases in the mean and standard deviation values of the annual emissivity between the contaminated ascending orbits data and the a priori RFI-free descending orbit ones. However, it is also apparent that some residual biases are still present after filtering in the ascending passes data, a priori induced by non eliminated residual RFI contaminations. However, they are now exhibiting a smaller temporal variability over the year on the order of the descending data ones. Apparently, this is because the grid node considered for that example is located in an area that is very often affected by RFI contaminations during ascending orbits at the particular incidence angle bin centered on 20°. A further processing step is therefore necessary to detect if the data at that particular angle can be used for salinity retrieval or is permanently contaminated and shall be therefore always discarded for SSS retrieval.

6.1.2.2. Multi-angular consistency check

A methodology to infer if an identified contaminated data subset at a given location, incidence angle and for a given orbit pass can be used for sss retrieval is to look at the multi-angular variability at that location.

As shown in the right plots of Figure ( ) below, the considered RFI signature shows a very different spatial pattern at 20° and 55°. The RFI induced variability zone located around a longitude of 40°W and a latitude of ~48°S at 20° incidence angle is displaced north westward at an incidence angle of 55° toward a center point at a longitude of ~44°W and a latitude of ~42.5°S. Therefore, there is apparently no RFI contamination at P1 (40°W,47.5°S) in the data at 55° incidence angle.

The varying multi-angular signature of RFI can therefore be used for outlier detection and removal. The evolution of the surface emissivities in several incidence angle bin along the latitudinal section S1 is shown in Figure ( ).
Figure: Example blow-up maps east of Argentina of the annual standard deviation of surface emissivities at an incidence angle of $\theta_{\text{bin}}^{2c} = 20^\circ$ (upper plots) and at an incidence angle of $\theta_{\text{bin}}^{9c} = 55^\circ$ (lower plots). Right plots are for descending orbits and left plots are for ascending ones.

As illustrated in Figure we shows latitudinal sections along S1 of the emissivities at several incidence angles and for both pass types, the annual mean emissivities at the latitude of point P1 ($47.5^\circ S$) are showing consistent mean values between ascending and descending passes over almost all incidence angles except at $20^\circ$ incidence. Looking farther North along S1, e.g., at $40^\circ N$, we see that the multi-angular/multi-orbit consistency changed. At that particular location, incidence angles from $45^\circ$ to $55^\circ$ now looks more affected by RFI (in term of mean biases) than at P1 and lower incidences including $20^\circ$ now seem consistent.
Figure: Annual mean emissivities in ascending orbits (thick curves) and descending orbits (thin curves) along the latitudinal section S1. The different colors indicate varying incidence angles values given in the legend. The blue point indicates the contaminated emissivity in descending passes at the latitude of P1.

A more compact view of the multi-angular/multi-orbit annual consistency in the data at P1 is given in Figure (), left plot. Here, we clearly see that the data at 20° in Ascending passes are not obeying the multi-look consistency over a one year period on average.

Figure: left: mean annual emissivities at location P1 as function of incidence angle ±1 std (ascending orbits=blue; descending orbits=red). Right: annual standard deviation of the surface emissivities at P1 as function of incidence angle ±1 (ascending orbits=blue; descending orbits=red). The thick black line indicate the mean standard deviation over all incidence angles from both passes. The thin dashed lines indicate the mean std ± three times the standard deviation over incidence angles of the annual standard deviation.
This is further

6.2 Ascending/Descending biases

6.3 Land contamination

6.4 Solar Radiation induced variability
7. Interpass Consistency and RFI mitigation