Atmospheric Correction Discussion

where we are today, and
where we might be in 2.5 years

B. Franz, PACE ST Meeting,
College Park, MD, January 2014
The goal of atmospheric correction (AC) is to convert observed top-of-atmosphere spectral radiance to water-leaving reflectance ($R_{rs}$) over the NUV-VIS spectral regime?
current NASA atmospheric correction approach

\[ L_t = \left( L_r + \left[ L_a + L_{ra} \right] + t_{dv} L_f + T_T v T_L g + t_{dv} L_w \right) t_{gv} t_{gs} f_p \]

\[ R_{rs} = \frac{L_w}{F_0 \cos(\theta_s) t_{ds} f_s} f_b f_\lambda \]
current NASA atmospheric correction approach

\[ L_t = (L_r + [L_a + L_{ra}] + t_{dv}L_f + T_sT_vL_g + t_{dv}L_w) t_{gv} t_{gs} f_p \]

\[ R_{rs} = \frac{L_w}{F_0 \cos(\theta_s) t_{ds} f_s f_b f_\lambda} \]

- Solar constant (irradiance, Thuillier 2003) & an adjustment for the Earth-Sun distance
- Instrument polarization correction
- Correction for out-of-band response
- Bidirectional reflectance correction
current NASA atmospheric correction approach

\[ L_t = \left( L_r + \left[ L_a + L_{ra} \right] + t_{dv} L_f + T_s T_v L_g + t_{dv} L_w \right) \left( t_{gv} t_{gs} f_p \right) \]

\[ R_{rs} = \frac{L_w}{F_0 \cos(\theta_s)} t_{ds} f_s f_b f_{\lambda} \]

not: no coupling between gaseous absorption and scattering terms
gaseous transmittance ($O_3$ & $NO_2$)
gases not directly considered (O₂ & H₂O)
current NASA atmospheric correction approach

\[ L_t = \left( L_r + \left[ L_a + L_{ra} \right] + t_{dv} L_f + T_s T_v L_g + t_{dv} L_w \right) t_{gv} t_{gs} f_p \]

\[ R_{rs} = \frac{L_w}{F_0 \cos(\theta_s) t_{ds} f_s} f_b f_{\lambda} \]
foam & whitecaps

$L_f(\lambda) = \rho_f(\lambda) \ [a \ (U_{10} + b)^3] \ F_0 \ cos(\theta_0) / \pi$

$\rho_f(\lambda) = \text{effective whitecap reflectance}$

22% from Koepke 1984
NIR spectral dependence from Frouin 1999

$U_{10} = \text{wind speed at 10-meters (\leq 12 m/s)}$

Monahan/Frouin/Koepke
Moore
Stramska
Stramska modified

Figure 8. Oceanic whitecap coverage as a function of wind speed. Different symbols are used for the developed wave field, the undeveloped wave field, and the decreasing wind speed. See text for details.
Sun glint

\[ L_g = F_0 L_{GN} \]

\[ T_s T_v = \exp \left[ - (\tau_r + \tau_a) \left( \frac{1}{\cos(\theta_0)} + \frac{1}{\cos(\theta)} \right) \right] \]

\( L_{GN} \) from Cox and Munk (1954)
- glint radiance normalized to no atmosphere & \( F_0 = 1 \)
- statistical function of windspeed
- flagged as high glint if \( L_{GN} > 0.005 \) and masked in Level-3

Actually using a two step iteration since we don’t know \( \tau_a \):

1. \( [L_t, \tau_a', W] \rightarrow L_t^{(1)} = L_t - TL_g \rightarrow \tau_a^{(1)} \)
2. \( [L_t^{(1)}, \tau_a^{(1)}, W] \rightarrow L_t^{(2)} = L_t^{(1)} - TL_g \rightarrow \tau_a^{(2)} \)

with initial guess of \( \tau_a' \sim 0.1 \)

Wang & Bailey 2001

Fig. 1. Normalized sun glint radiance \( L_{GN} \) as a function of the sensor-viewing angle (solar zenith angle, 40°) and for (a) various relative azimuthal angles with surface wind speed of 5 m/s and (b) various surface wind speeds with a relative azimuthal angle of 20°.
molecular (Rayleigh) scattering

Using pre-computed look-up tables from Ahmad & Fraser vector radiative transfer simulations for wind-roughened ocean surface.

Input bandpass-integrated Rayleigh optical thickness was computed using the model of Bodaine et al. (1999).

Rayleigh radiances (with polarization, I, Q, U) are retrieved from look up tables and adjusted given:

- solar & satellite viewing geometries \((\theta_0, \theta, \Delta\phi)\)
- windspeed (a proxy for surface roughness)
- atmospheric pressure (to adjust for change in optical thickness, \(\tau_r\))

Rayleigh reflectance calculable to \(~0.2\%\) (bias)
  based on RT intercomparisons, before vicarious calibration
current NASA atmospheric correction approach

\[ L_t = \left( L_r + [ L_a + L_{ra} ] + t_{dv} L_f + T_s T_v L_g + t_{dv} L_w \right) t_{gv} t_{gs} f_p \]

\[ R_{rs} = \frac{L_w}{F_0 \cos(\theta_s) t_{ds} f_s f_b f_{\lambda}} \]
aerosol contribution (basic concept)

assume $L_w(\lambda) = 0$ at two NIR (or SWIR) bands, or that it can be estimated with sufficient accuracy.

retrieve aerosol reflectance in each NIR band as

$$[L_a + L_{ra}] = \frac{L_t}{t_{gv} t_{gs} f_p} - (L_r + t_{dv} L_f + T_s T_v L_g + t_{dv} L_w)$$

$$\rho_a = [L_a + L_{ra}] \frac{\pi}{F_0 \cos(\theta_0)}$$

use spectral dependence of retrieved NIR aerosol reflectance ($\varepsilon$) to select the most appropriate aerosol model from a suite of pre-computed models

use NIR aerosol reflectance and selected aerosol model to extrapolate aerosol reflectance to visible wavelengths
we estimate $L_w(NIR)$ using a bio-optical model

1) convert $L_w(670)$ to $b_b/(a+b_b)$ via Morel f/Q and retrieved Chl$_a$

2) estimate $a(670) = a_w(670) + a_{pg}(670)$ via NOMAD empirical relationship

$$a(670) = e^{(\ln(C_a)*0.9389-3.7589)} + a_w(670)$$

3) estimate $b_{bp}(NIR) = b_{bp}(670) (\lambda/670)^\eta$
via Lee et al. 2002

$$\eta = 2.0 \times [1. - 1.2 \times e^{(-0.9*R_s(443)/R_s(555))}]$$

4) assume $a(NIR) = a_w(NIR)$

5) estimate $L_w(NIR)$ from $b_b/(a+b_b)$ via Morel f/Q and retrieved Chl$_a$

Bailey et al., Optics Express, 2010
locations (white) where $L_w$(NIR) is significant

locations of application of bio-optical model

not applied when Chl < 0.3 mg m$^{-3}$
weighted application when 0.3 < Chl < 0.7 mg m$^{-3}$
fully applied when Chl > 0.7 mg m$^{-3}$

Bailey et al., Optics Express, 2010
aerosol model tables

- vector RT code (Ahmad-Fraser)
- based on AERONET size distributions & albedos
- 80 models (10 size fractions within 8 humidities)
  - 100% coarse mode to 95% fine mode
  - non- or weakly absorbing
- LUT: extinction, albedo, phase function, ss->ms, \( t_d \)
  - function of path geometry (or scattering angle)
- model selection discriminated by relative humidity

Typical Size Distributions

<table>
<thead>
<tr>
<th>Rh</th>
<th>( r_{vf} )</th>
<th>( \sigma_f )</th>
<th>( r_{vc} )</th>
<th>( \sigma_c )</th>
<th>( r_{vf}/r_{oVF} )</th>
<th>( r_{vc}/r_{ocv} )</th>
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<td>0.30</td>
<td>0.150</td>
<td>0.437</td>
<td>2.441</td>
<td>0.672</td>
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<td>0.50</td>
<td>0.152</td>
<td>0.437</td>
<td>2.477</td>
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<td>0.85</td>
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<td>0.90</td>
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<td>4.638</td>
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<td>0.95</td>
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<td>5.549</td>
<td>0.672</td>
<td>1.648</td>
<td>2.293</td>
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select the two sets of 10 models (10 size fractions) with relative humidity (RH) that bound the RH of the observation.

find the two models that bound the observed epsilon within each RH model family.

\[ \varepsilon^{obs}(748,869) = \frac{\rho_a(748)}{\rho_a(869)} \rightarrow \varepsilon^{mod}(748,869) \]

use model epsilon to extrapolate to visible.

\[ \rho_a(\lambda) = \rho_a(869)\varepsilon^{mod}(\lambda,869) \]

compute weighted average, \( \bar{\rho}_a \), between models within each RH family, and then again between bounding RH solutions.

\[ \left[ L_a + L_{ra} \right] = \bar{\rho}_a(\lambda) \frac{F_0 \cos(\theta_0)}{\pi} \]

*actually done in single scattering space and transformed to multi-scattering*
can’t distinguish absorbing aerosols using NIR spectral dependence alone

\[ \alpha = 0.02 \quad \alpha = 0.14 \quad \alpha = 0.40 \]
Measurements in two spectral bands in the red, near-infrared, and/or short-wave infrared are not sufficient, in the general case, to determine the perturbing effects of the atmosphere and surface in the visible.

No sensitivity to aerosol absorption, information about aerosol altitude needed.

Absorption effect as a function of wavelength (top left), air mass (bottom left), and aerosol pressure level (top right) for continental, urban, desert dust, and biomass burning aerosol models. The effect increases in magnitude with decreasing wavelength, decreasing aerosol pressure level, and increasing air mass.

Frouin
note: NIR aerosol reflectance is an error bucket

- any error in subtraction of Rayleigh, whitecaps, or glint, or any signal that is not identified and subtracted (e.g., thin cirrus, cloud edges, straylight, atmospheric adjacency) will add to the aerosol reflectance in the NIR and be extrapolated to the visible via the aerosol model.

- many of these sources are approximately white, so likely the effect is to flatten the retrieved spectral dependence proportionate to the residual signal.

- both the aerosol concentration and aerosol type will be altered by such errors, and thus the aerosol properties will be inaccurate, but the reflectance that is subtracted “may” be approximately correct.
current NASA atmospheric correction approach

\[ L_t = \left( L_r + \left[ L_a + L_{ra} \right] + t_{dv} L_f + T_s T_v L_g + t_{dv} L_w \right) t_{gv} t_{gs} f_p \]

\[ R_{rs} = \frac{L_w}{F_0 \cos(\theta_s) t_{ds} f_s} f_b f_\lambda \]

bidirectional reflectance correction
brdf correction

to account for shape of sub-surface light-field due to position of the Sun and optical properties of the water column.


given radiant path geometry \((\theta_0, \theta, \Delta \varphi)\), windspeed \((w)\) and \textbf{Chl}

\[
\text{Chl} = f(R'_\text{rs}(\lambda))
\]

\[
R'_\text{rs} = \frac{L_w}{F_0 \cos(\theta_s) t_{ds} f_s} f_\lambda
\]

\[
f_b(\lambda, \theta_0, \theta, \Delta \varphi, \text{Chl}, w) = (R_0 f_0/Q_0) / (R f/Q)
\]

\[
f/Q \text{ relates subsurface irradiance reflectance to radiance reflectance}
\]

\[
R \text{ includes all reflection/refraction effects of the air-sea interface}
\]

\[
R_{\text{rs}}(\lambda) = R'_\text{rs}(\lambda) f_b(\lambda, \theta_0, \theta, \Delta \varphi, \text{Chl}, w)
\]

\[
\text{Chl} = f(R_{\text{rs}}(\lambda))
\]
over ocean, we attempt to process all pixels with valid radiometry and navigation, that are not classified as cloud.

the standard cloud mask is just a threshold on surface + aerosol reflectance (excluding glint) at ~865nm.

\[
\begin{align*}
\left[L_a + L_{ra}\right] + t_{dv}L_f + t_{dv}L_w &= \frac{L_t}{t_{gv}t_{gs}} - \left(L_r + T_sT_vL_g\right) \\
\rho_c &= \left[\frac{L_t}{t_{gv}t_{gs}} - \left(L_r + T_sT_vL_g\right)\right] \frac{\pi}{F_0 \cos(\theta_0)} \\
\rho_c &> 0.022 = \text{cloud}
\end{align*}
\]

note: any bright signal in NIR will be classified as cloud (e.g., uncorrected glint, ice, high suspended sediment loads, bottom reflectance, high concentration of coccolithophores).
summary of current atmospheric correction

• correction for atmospheric absorption
  – O3, NO2  (O2 & H2O only for out-of-band responsivity corrections)
  – using coincident O3 and climatological NO2

• surface and sub-surface corrections (windspeed dependent)
  – Sun glint, statistical model based on Cox & Munk
  – brdf (fresnel + Morel f/Q)

• subtraction of radiance scattered by air molecules
  – pre-computed Rayleigh scattering look-up tables w/polarization
  – function of geometry and windspeed, also uses surface pressure

• subtraction of radiance scattered by aerosols + Rayleigh-aerosol
  – aerosol contribution derived from reflectance in 2 NIR bands + models
  – aerosol models derived from AERONET measurements (non- or weakly-absorbing marine and coastal aerosols)
  – depends on ancillary relative humidity
### summary of ancillary data requirements

<table>
<thead>
<tr>
<th>ancillary data</th>
<th>ancillary source</th>
<th>uses</th>
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<tbody>
<tr>
<td>atmospheric pressure</td>
<td>NCEP</td>
<td>Rayleigh</td>
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<tr>
<td>relative humidity</td>
<td>NCEP</td>
<td>aerosol models</td>
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<td>wind speed</td>
<td>NCEP</td>
<td>white caps, Sun glint, Rayleigh</td>
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<td>ozone</td>
<td>OMI/TOMS</td>
<td>transmittance</td>
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<tr>
<td>NO2</td>
<td>Sciamachy/OMI/GOME</td>
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<td>f/Q (bidirectional reflectance distributions)</td>
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</tbody>
</table>
how well does it work?
MODISA Rrs Validation (SeaBASS + AERONET-OC)

Rrs(443)  Rrs(488)  Rrs(547)

Mean APD 12-13%, Mean Bias < 10%, $R^2 > 0.9$
**how well does it work?**

MODISA Rrs Validation (SeaBASS + AERONET-OC)

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The linear regression algorithm has been changed to reduced major axis.
Rrs uncertainty goals (PACE SDT)

*open ocean, clear-water, marine aerosols*

\[
[\rho_w(\lambda)]_N \quad \lambda=400-710\text{nm}, \text{ maximum of 0.001 or 5\% (VIS)}
\]
\[
[\rho_w(\lambda)]_N \quad \lambda=350-400\text{nm}, \text{ maximum of 0.002 or 10\% (NUV)}
\]

in terms of \( R_{rs}(\lambda) = [\rho_w(\lambda)]_N/\pi \), that is:

\[
\Delta R_{rs}(\lambda) = 3e^{-4} \text{ (sr}^{-1}, \text{ VIS)}
\]
\[
\Delta R_{rs}(\lambda) = 6e^{-4} \text{ (sr}^{-1}, \text{ NUV)}
\]
how well does it work?
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PACE SDT Goal for Rrs(VIS)
\[ \Delta R_{rs}(VIS) = 3 \times 10^{-4} \text{ sr}^{-1} \text{ or } 5\% \]

Current Approach
\[ \Delta R_{rs}(VIS) \sim 1 \times 10^{-3} \text{ sr}^{-1} \text{ or } 12\% \ (22\% \ 412) \]

goal is factor of 3 reduction ... seems achievable!
questions on heritage algorithm?
where do we go from here?

I see two complementary paths:

1. adapt and improve the heritage algorithm to support PACE SDT notional instruments (*Franz, Gao*)

2. develop a completely different approach (i.e., simultaneous retrieval of atmosphere and $R_{rs}$/IOPs (*Chowdhary, Frouin*))

with both paths potentially benefiting from work within the Team, e.g.:

- absorbing aerosol identification (NUV) and modeling
- aerosol height (O2 A-band)
- absorbing gas corrections (H2O)
- whitecaps
- cloud flagging or cloud corrections
Implementation
NASA Standard Processing Code

Level-1 to Level-2
(common algorithms)

SeaWiFS L1A
MODISA L1B
MODIST L1B
OCTS L1A
MOS L1B
OSMI L1A
CZCS L1A
MERIS L1B
OCM-1 L1B
OCM-2 L1B
VIIRS L1A
GOCI L1B
L8 OLI L1T

Level-2 Scene

ancillary data

sensor-specific tables: Rayleigh, aerosol, etc.
in situ water-leaving radiances (MOBY)

Level-2 to Level-3

vicarious calibration gain factors

Level-3 Global Product

predicted at-sensor radiances

water-leaving reflectances & derived prods

observed radiances
l2gen already supports many different atmospheric correction options and algorithms

• aerosol model selection
  – specifiable band pairs, including NIR-SWIR switching
  – specifiable model suites, with or without RH stratification
  – model selection in single or multi-scattering space
  – various methods of NIR Lw estimation (Bailey et al. 2010, Ruddick et al. 2000)

• simultaneous atmospheric correction and IOP retrieval
  – Spectral Matching Algorithm (Gordon et al. 1997, Moulin et al. 2001)

• modular components for whitecaps, glint, gas transmittances, etc.
where do we go from here?

I see two complementary paths:

1. adapt and improve the heritage algorithm to support PACE SDT notional instruments \((Franz, Gao)\)

2. develop a completely different approach (i.e., simultaneous retrieval of atmosphere and \(R_{rs}/IOPs\) \((Chowhdary, Frouin)\)

with both paths potentially benefiting from work within the Team, e.g.:

- absorbing aerosol identification (NUV) and modeling
- aerosol height (O2 A-band)
- absorbing gas corrections (H2O)
- whitecaps
- cloud flagging or cloud corrections

using l2gen as the implementation framework
discussion
Is it reasonable to expect a working implementation of the heritage algorithm for a PACE-like radiometer within 2.5 years?

yes
1. modify l2gen to support hyperspectral (in progress)
2. add simple water-vapor correction (working with Gao)
3. start testing on HICO, AVIRIS

4. start developing/testing algorithm enhancements (e.g.)
   – aerosol selection in multi-scattering space
   – use of more than two bands (minimization over NIR-SWIR atm. windows), adaptive NIR-SWIR band-set selection
   – incorporating other developments within ST (abs. aerosol detection, whitecaps, cirrus, etc.)
Is it reasonable to expect a working implementation of at least one alternative algorithm (e.g., ACROSS) within 2.5 years?
How will we evaluate algorithm performance/behavior?

– simulated data: controlled experiment, answer is known, but does not test real-world conditions, may favor one algorithm where forward and inverse models are common

– aircraft/spacecraft data + co-incident field measurements: no perfect match to PACE notional sensors (OCI, OCI+, etc.), sensor-specific calibration issues may (will) confound results, uncertainty in field measurements
What are the likely advancements we can demonstrate in 2.5 years?

- retrieval of Rrs in NUV-VIS for open ocean, marine aerosols?

- absorbing aerosol detection? correction? accounting for aerosol height?

- can we improve identification and correction for non or weakly-absorbing aerosols?

- improved cloud detection (cloud correction)?

- improved glint correction? whitecap correction?

- improved brdf? should we even apply brdf before IOP inversion?
What are the major challenges?

- Rayleigh-aerosol interaction in NUV, sensitivity to error in aerosol absorption

- coupling of absorbing aerosols and CDOM in NUV

- accurate correction for absorbing gases over NUV-SWIR (water vapor)

- is solar irradiance knowledge sufficient (Thuillier 2003)?
How can polarimeter measurements contribute to atmospheric correction, and what kind of polarimeter is required?

- is multi-angle required
- what spectral bands?
- co-registered to the radiometer swath?
- co-registered to radiometer spatial sampling resolution?
Can we reduce dependency on ancillary sources?
Where are the gaps (known issues, not being worked)?
Frouin: “atmospheric correction should”

current NASA heritage algorithm

1. work in turbid, optically complex waters maybe, if NIR bio-optical model is valid
2. work in the presence of moderate sun glint maybe, if windspeed is accurate
3. work in the presence of semi-transparent clouds maybe, will be treated as aerosol
4. work in the presence of whitecaps yes, if windspeed is accurate
5. work when air mass is large how large? plane-parallel assumed
6. handle adjacency effects no, only instrument straylight, no atmospheric adjacency
   – due to proximity of clouds, ice, land, etc.
7. handle situations of absorbing aerosols no, spectral dep in NIR is not unique
8. provide per-pixel uncertainties on Rrs no, not directly
9. be insensitive to radiometric calibration errors yes, with uncertainty of order 1e-3 sr\(^{-1}\) based on match-ups

and, work in open waters, in the presence of non-absorbing, low concentration aerosols, no adjacency effects, no whitecaps, no glint, no spherical effects
back-up
For dust ($\omega=0.878$) & $\tau_a=0.1$, a 1-km error in aerosol layer height corresponds to 0.3% difference in $L_t$. This translates into a 3% difference in $L_w$. The error increases with increasing $\tau_a$.

For an aerosol layer at 3-km & $\tau_a=0.1$, a change from $\omega=0.878$ to $\omega=0.918$ corresponds to 1% difference in $L_t$. This translates into a 10% difference in $L_w$. The error increases with increasing $\tau_a$.
Frac. Aer. Refl. = \[ \frac{\text{Ref (Rayl+Aer.)}}{\text{Ref (Rayl.)}} - 1 \]  

\[ \theta_o = 42^\circ \]  
\[ \theta = 0^\circ \text{ (Nadir)} \]