Influence of light absorbing particles on snow albedo and radiation

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June 25, 2012
Alpine Summer School: Climate, Aerosols and the Cryosphere
Valsavarenche, Italy
1. Background on Snow Darkening

2. Modeling
   - Background
   - Monte Carlo Modeling
   - Scattering Phase Function

3. Impurities
   - Grain Size and Aging
   - Impurity Mixing State
“Pure as snow”

“Be thou as chaste as ice, as pure as snow, thou shalt not escape calumny” — Shakespeare, Hamlet.
The color of snow

- Landsat Thematic Mapper imagery
- From Jeff Dozier (UCSB), “Mountain hydrology, snow color, and the fourth paradigm”
The color of snow and ice

(a) Large variability in near-infrared albedo of snow with grain size

(b) Measured spectral reflectance of different surfaces on Greenland (Bøggild et al, 2010)
Snow albedo perturbation from black carbon (soot)

(c) Influence of ~1 gram of black carbon distributed over a square meter of snow (Niwot Ridge, CO)

(d) 3 days later the snow column was depressed

- Background impurity?
Long-term change in BC deposition on Greenland

- BC concentrations in ice at D4, Greenland (McConnell et al, 2007)
- BC peak near ~1910
- Biomass burning contribution indicated with vanillic acid
Dust in the San Juan Mountains

MODIS imagery

NASA Earthobservatory Image of the Day
(g) Neff et al, 2008: 4-fold increase in dust deposition
(h) Painter et al, 2010: Earlier date of 90% snow ablation from dust
(i) Dust-induced change in hydrograph at Lees Ferry
Snow darkening from ash deposition

http://www.sonnblick.net/portal/content/view/191/248/lang,de/

Visible snow darkening in Austria, attributed to ash from Icelandic eruption (Eyjafjallajökull) several thousand kilometers away
Motivation: Radiative forcing (IPCC AR4)

- **RF Terms**
  - **Long-lived greenhouse gases**
    - CO$_2$
    - N$_2$O
    - CH$_4$
    - Halocarbons
  - **Ozone**
    - Stratospheric
    - Tropospheric
  - **Stratospheric water vapour from CH$_4$**
  - **Surface albedo**
  - **Land use**
  - **Direct effect**
  - **Total Aerosol**
    - Cloud albedo effect
  - **Linear contrails**
  - **Solar irradiance**
  - **Total net anthropogenic**

- **RF values (W m$^{-2}$)**
  - CO$_2$: 1.66 [1.49 to 1.83]
  - N$_2$O: 0.48 [0.43 to 0.53]
  - CH$_4$: 0.16 [0.14 to 0.18]
  - Halocarbons: 0.34 [0.31 to 0.37]
  - Stratospheric water vapour from CH$_4$: 0.07 [0.02 to 0.12]
  - Surface albedo: -0.2 [-0.4 to 0.0]
  - Land use: 0.1 [0.0 to 0.2]
  - Direct effect: -0.5 [-0.9 to -0.1]
  - Cloud albedo effect: -0.7 [-1.8 to -0.3]
  - Linear contrails: 0.01 [0.003 to 0.03]
  - Solar irradiance: 0.12 [0.06 to 0.30]
  - Total net anthropogenic: 1.6 [0.6 to 2.4]

- **Spatial scale**
  - Global
  - Continental to global
  - Local to continental

- **LOSU**
  - High
  - Med
  - Low
Why do ppb levels of soot perturb albedo?

Part-per-billion levels of BC significantly reduce snow albedo because:

- Black carbon visible absorptivity is $\sim 10^5$ greater than ice
- Snow scatters visible radiation efficiently via refraction
  - A typical reflected blue photon undergoes $\sim 3000$ scattering events before emerging from the top of snowpack
- Longer persistence in near-surface snow than atmosphere.
A Model for the Spectral Albedo of Snow. I: Pure Snow

WARREN J. WISCHEMCE AND STEPHEN G. WARREN

National Center for Atmospheric Research, Boulder, CO 80307

(Manuscript received 15 April 1980, in revised form 28 August 1980)

- Mie Theory applied to derive optical properties of ice particles and impurities
- Multiple-scattering approximation (delta-Eddington) to represent transfer in the bulk medium
- Pure-snow model over-predicted reflectance in the visible spectrum
A Model for the Spectral Albedo of Snow. II: Snow Containing Atmospheric Aerosols

Stephen G. Warren¹ and Warren J. Wiscombe

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(Manuscript received 15 April 1980, in final form 28 August 1980)

- Warren and Wiscombe (1980), J. Climate
- Minute concentrations of soot reduced model-measurement bias
What does snow albedo depend on?

- Snow grain size
- Solar zenith angle
- Content of absorbing impurities like black carbon, dust, algae, ash
- Snow thickness and underlying ground albedo

More nuanced definitions of albedo:

- *Spectral reflectance*: The reflectance at a particular wavelength or frequency
- *Broadband albedo*: Integrated reflectance over a spectral range ($\lambda_1 - \lambda_2$). Depends on spectral distribution of incident light:

\[
A_{bb} = \frac{\int_{\lambda_1}^{\lambda_2} r_{\lambda} F_{\lambda}^{\downarrow} d\lambda}{\int_{\lambda_1}^{\lambda_2} F_{\lambda}^{\downarrow} d\lambda} = \frac{F_{bb}^{\uparrow}}{F_{bb}^{\downarrow}} \quad (1)
\]

- *Directional reflectance*: Reflectance into a given direction, given a direction of incidence
- *Hemispheric albedo*: Hemispherically-integrated reflectance
- *Hemispheric broadband albedo*: ??
Directional reflectance: The BRDF

- Function describing directional reflectance, given angle of incidence, is the *bidirectional reflectance distribution function* (BRDF): \( \rho_{\lambda}(\theta_i, \phi_i; \theta_r, \phi_r) \)
- Left: normalized BRDF of snow, measured in different parts of the spectrum (Aoki *et al.*, 2000)
  - Normalization is relative to that measured in nadir direction (center of plot)
  - Radial distance from center is viewing zenith angle (\( \theta_r \))
  - Radial angle is viewing azimuthal angle (\( \phi_r \))
  - Bottom(ish) is forward scattering direction
Basic ingredients for modeling snow radiative transfer

1. Knowledge of how likely the element is to interact with (i.e., scatter or absorb) incident radiation: *mass extinction coefficient* \((k_{\text{ext}})\)

2. Knowledge of the relative likelihood that the element will scatter or absorb incident light: *single-scatter albedo*:

\[
\tilde{\omega} = \frac{k_{\text{sca}}}{k_{\text{ext}}}
\]  

3. Mathematical description of how the element (snow grains) scatters, or re-directs radiation at different wavelengths: *scattering phase function*

4. The mass or volume density of the element (snow grains)

5. A model of radiative transfer that accommodates *multiple scattering*

6. For inclusion of impurities like black carbon and dust, we also need to know properties 1–4 for these “elements”
Building an analytical RT model

- Complicated!
- Start with the Extinction Law: \( I(\tau) = I_0 e^{-\tau} \)
- Extinction optical depth \( \tau = \beta_e ds = k_{\text{ext}} L \)
- \( dl = dl_{\text{ext}} + dl_{\text{emit}} + dl_{\text{scat}} \)
- What are contributions to \( dl_{\text{scat}} \)?
  1. The scattering coefficient (\( \beta_s \))
  2. Radiation traveling in any direction (\( \hat{\Omega} \)) that gets scattered into the direction of interest (\( \hat{\Omega}' \))
- We can characterize factor (2) with the scattering phase function: \( p(\hat{\Omega}', \hat{\Omega}) \)
- If scattering occurs within our infinitesimal volume, \( p \) characterizes the probability that the photon will be scattered into the direction \( \hat{\Omega} \) (given \( \hat{\Omega}' \)).
- The actual probability is \( p/4\pi \), so that:

\[
\frac{1}{4\pi} \int_{4\pi} p(\hat{\Omega}', \hat{\Omega}) d\omega' = 1 \tag{3}
\]
Multiple scattering: One solution for snow

\[
Q\alpha_d = 2P \left[ (1 - \gamma + \omega^*b^*)(1 - \tau_0^*) - \frac{\gamma\omega^*(1 + b^*)}{1 - \omega^*} \right] \exp(-\tau_0^*) - 2P \left[ \omega^*(1 + b^*) \left( \frac{2}{\xi^2} + \frac{\gamma\tau_0^*}{1 - \omega^*} \right) \right.
\]
\[
\left. + (1 - \gamma + \omega^*b^*)\tau_0^{*2} \right] \text{Ei}(-\tau_0^*) + \frac{2\omega^*(1 + b^*)}{\xi^2} \left[ Q^+\{\text{Ei}[-(1 + \xi)\tau_0^*] + \xi - \ln(1 + \xi) \right]
\]
\[
- Q^-\{\text{Ei}[-(1 - \xi)\tau_0^*] - \xi - \ln|1 - \xi| \} \right] - \omega^*b^* (Q^+ - Q^-),
\]

- *Wiscombe and Warren* (1980): Hemispheric albedo of a snowpack of finite thickness at a single wavelength for diffuse incident light (e.g., under a thick cloud)
- Again, the challenge is accounting for *multiple scattering*
Monte Carlo modeling

- Analytical approximations are possible when numerous assumptions are made.
- An alternative approach is with *Monte Carlo* modeling: repeated random sampling from probability density functions to obtain statistical representation of the system’s behavior.
- This technique is applied in numerous fields (not just radiative transfer modeling).
**Pros:**

- Given enough sampling, the technique is highly accurate. (Accuracy limited much more by uncertainty in PDFs than in computational technique)
- It is robust: 3-D modeling and heterogeneous geometries can be more easily incorporated. Not restricted to plane-parallel assumptions
- Considering radiative transfer from the discrete photon perspective, this approach is physically-based and intuitive
- Relatively easy to program
Monte Carlo modeling

Cons:

- Computationally expensive – impractically so in many cases. Need to simulate many photons to obtain reliable results, and the number needed can become unreasonably large, especially with optically-thick media.
- Photons can become “lost” deep inside optically thick media like clouds and snow, requiring excessive computational time to resolve.
- Cannot capture anywhere near the true number of photons! Rather, we hope to capture a representative sample.
Monte Carlo “decision tree” for modeling photons

1. Launch photon
2. Determine path length to next event
3. Move photon
4. Record where it exited. DONE.
5. If photon out of domain?
   a. Yes
      - Add to Heating rate. DONE.
   b. No
      - Scattered or absorbed?
        a. Scattered
           - Determine new direction
        b. Absorbed
           - Add to Heating rate. DONE.
Again, we can think of $p$ (or $p/4\pi$) as a probability density function. It must integrate to 1. (If a photon is scattered, it must go somewhere. And, we cannot end up with more photons, through scattering, than we started with).

What controls $p$?
- Particle size
- Particle shape

We can greatly simplify when particles are spheres (or when we assume particles are spheres). In this case, the geometry is isotropic, and we only need to consider the scattering phase angle: $\Theta$ (instead of four directions in the general case):

$$\cos \Theta = \hat{\Omega}' \cdot \hat{\Omega}$$ (4)

$\Theta$ is simply the angle between the incident and scattered photon

$$\frac{1}{2} \int_{-1}^{1} p(\cos \Theta) \, d \cos \Theta = 1$$ (5)
Isotropic scattering

- Simplest scattering case: *isotropic scattering*: equal scattering in all directions. No information about scattering direction from incident direction.

\[ p(\cos \Theta) = 1 \]  

(6)

- Isotropic scattering example inside of a cloud: “Aimless wandering”. In this case, photon emerges from top of snowpack/cloud and contributes to albedo, but in others it would have passed through, contributing to *diffuse transmittance*. Almost no *direct transmittance*. 
Photon outcomes

Four possibilities for a photon incident on top of snowpack:

1. Passes through snow without being scattered once, or absorbed. The fraction that experiences this is direct transmittance ($t_{\text{dir}}$).

2. Scattered one or more times and then emerge from the bottom of the snowpack (where it is likely to be absorbed by ground). This fraction is the diffuse transmittance ($t_{\text{dif}}$).

3. Scattered one or more times and then emerge from the top of the snow. This fraction represents the reflectance (or albedo).

4. It can be absorbed (by ice, impurities, or air). This fraction is the absorptance.

5. The four fractions (probabilities) must sum to 1 for conservation:

$$t_{\text{dir}} + t_{\text{dif}} + r + a = 1$$  (7)
The asymmetry parameter

- Scattering by real particles (snow grains) is never isotropic.
- To accurately model scattering, we might need a complicated function to describe \( p(\hat{\Omega}', \hat{\Omega}) \).
- **For spheres, this function can be computed analytically with Mie Theory.**
- But in many cases, we may only be concerned with flux (not intensity), in which case it is sufficient to know the relative proportion of photons scattered in the forward and backward directions (i.e., the backscatter fraction), or the mean scattering angle.
- The asymmetry parameter \((g)\) does this, describing the average value of \( \cos \Theta \) for a large number of scattered photons:

\[
g = \frac{1}{4\pi} \int_{4\pi} p(\cos \Theta) \cos \Theta d\omega \quad (8)
\]
The asymmetry parameter

- Range of asymmetry parameter:

\[-1 \leq g \leq 1\]  \hspace{1cm} (9)

- If $g > 0$, photons are preferentially scattered into the forward hemisphere
- If $g < 0$, photons are preferentially scattered into the backward hemisphere
- In the case of isotropic scattering, what is $g$?
- What would $g = -1$ imply?
- Can $g = 0$ for non-isotropic scattering?
- Aerosols, snow grains, and cloud droplets that scatter visible radiation typically have: $0.8 \leq g \leq 0.9$, meaning they are strongly forward-scattering
Scattering and the asymmetry parameter

- *Isotropic* scattering on left, *forward-scattering* on right
- Cloud/snow optical depth is the same in both cases
- In $g = 0.85$ case, photons are much less likely to undergo rapid redirection
- Which case would exhibit greater *diffuse transmittance*? And greater albedo?
Everyday experience: *Rayleigh scattering* from air molecules (left) is **not** forward scattering. Haze or atmospheric aerosols (right) scatters in the forward direction.
More examples of cloud/snow scattering

- The random paths of 100 photons
- All of these cases have conservative scattering \(\tilde{\omega} = 1\), but with different values of \(g\)
- Incident photon angle is \(\theta = 30^\circ\)
Often $g$ is sufficient to describe scattering direction, but sometimes we need more detail. The Henyey–Greenstein function “fills-in” a full scattering phase function, using only $g$.

$$
\rho_{HG}(\cos \Theta) = \frac{1 - g^2}{(1 + g^2 - 2g \cos \Theta)^{3/2}}
$$

Convenient and reasonably accurate in many cases, but it is an example of downscaling (creating higher resolution data from low-resolution input).
Henyey–Greenstein phase function

For positive $g$, the function peaks increasingly in the forward direction (as we would hope), but remains smooth. Thus, it works fairly well for snow grains, cloud droplets and aerosols, especially if they are spherical.
Monte Carlo model demonstration
Improved modeling of snow radiative transfer

- Hyperspectral two-stream multiple scattering approximation (Toon et al, 1989) to solve for fluxes throughout the snow column
- Multiple layers account for vertical inhomogeneity
- Two streams: appropriate for energy-balance and climate modeling, but not for remote sensing applications
- Visible radiation penetrates deeply in snow, whereas near-IR energy is absorbed in top ∼ 1 mm of snowpack

**Figure:** Flanner and Zender (2005), GRL
Modeled and measured spectral albedo

A single layer with $r_e = 100\,\mu m$

Figure: Data from Grenfell et al. (1994), JGR
Modeled and measured spectral albedo

Figure: Data from Grenfell et al. (1994), JGR

Layer 1: $r_e = 45 \mu m$
Layer 2: $r_e = 100 \mu m$
Albedo perturbation from impurities
Albedo perturbation from impurities
Albedo perturbation from impurities
Albedo perturbation from impurities
Albedo perturbation from impurities
A perfect (spectral) match

The spectral range of peak albedo reduction matches that of peak irradiance. Surface irradiance shifts, however, with clouds, solar zenith angle, and absorbing gases.
Albedo perturbation from impurities

Simulate it yourself at: http://snow.engin.umich.edu
Modeled and measured snow reflectance

- First snow doping experiments with known mixing ratios of black carbon
- Good agreement between measured and simulated reflectance offers confidence that radiative transfer models can be applied to study snow darkening

*Figure:* Experiments conducted by Charlie Zender and Florent Domine, LGGE, France
Modeled and measured snow reflectance

Figure: Spectrally weighted snow albedo over the 300–500 nm solar spectrum: derived from lab experiments (dots, ±1 standard deviation) and modelled using SNICAR (shaded bands). Hadley and Kirchstetter (2012), Nature Climate Change
The importance of snow grain size

- Snow exhibits large variability in grain size ($30 < r_e < 2000 \mu m$)
- Grain size determines pure snow albedo, depth profile of absorption, and the magnitude of perturbation by impurities
The importance of snow grain size

- Snow exhibits large variability in grain size \((30 < r_e < 2000 \mu\text{m})\)
- Grain size determines pure snow albedo, depth profile of absorption, and the **magnitude of perturbation by impurities**
Why the dependence on grain size?

- Snow density: 350 kg m\(^{-3}\)
- Zenith angle: 60°
- Wavelength: 550 nm

Radiative intensity diminishes more slowly with depth in snow with larger \(r_e\) because:
  - It is less extinctive (per unit mass)
  - It scatters more strongly in the forward direction

Therefore, photons travel through a greater optical depth of (homogeneously interspersed) BC
Concurrent darkening from dust and organics

- The contribution of a given impurity to snow darkening is diminished in the presence of other impurities. (true for all RF agents).

- Globally, BC/snow forcing is reduced ~ 20% when dust is included
What happens when BC is *inside* the snow grain

- Different optical representations
- DEMA is likely most appropriate because **there are typically more BC particles than snow grains**. How?
  - BC particle radius: $\sim 20$ nm
  - Snow grain radius: $\sim 200$ μm
  - Thus, typical snow grain is $\sim 12$ orders of magnitude larger (by volume) than BC. Typical BC mass concentration?
Absorption enhancement from BC in ice

Enhancement depends on particle size (Flanner et al., 2012, ACP)
Absorption enhancement from BC in ice

- For typical BC and ice particle sizes and BC volume fractions that we might expect for snow, the absorption enhancement is $\sim 1.9$ (Flanner et al., 2012).
- The “screening effect” occurs when too much BC is packed into a snow grain (Chylek et al., 1983, previous slide).
Global aerosol models simulate most BC deposition to snow occurring within precipitating hydrometeors (*Flanner et al.*, 2012, ACP)