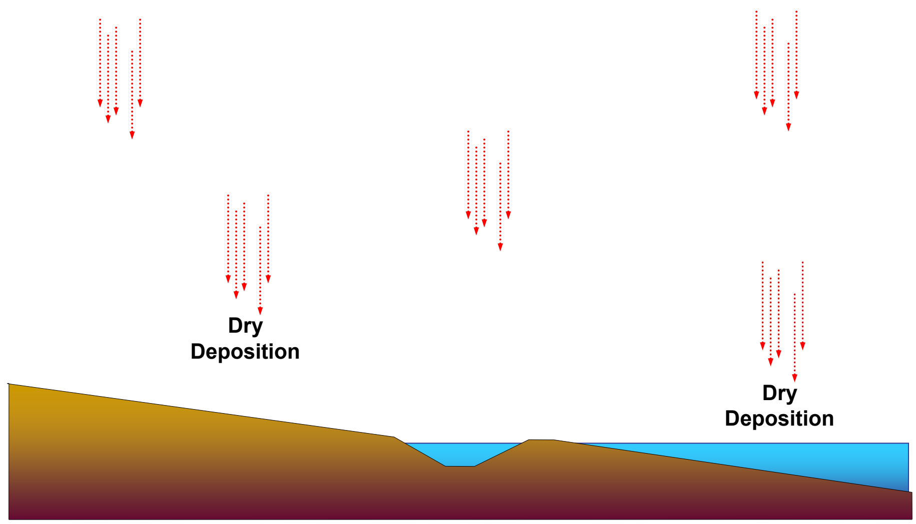
Lecture 8: Dry Deposition



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Dry Deposition

Dry deposition – removal of gases and particles by a direct transfer from the atmosphere to the surface (in the absence of precipitation)**.**

Three separate steps

(1) Species must be transported close to the surface

(2) Species must cross to the surface

(3) Species must be taken up on the surface.

The factors that govern the dry deposition of a gaseous species or a particle are the level of atmospheric turbulence, the chemical properties of the depositing species, and the nature of the surface itself. The level of turbulence in the atmosphere, especially in the layer nearest the ground, governs the rate at which species are delivered down to the surface. For gases, solubility and chemical reactivity may affect uptake at the surface. For particles, size, density, and shape may determine whether capture by the surface occurs. The surface itself is a factor in dry deposition. A nonreactive surface may not permit absorption or adsorption of certain gases; a smooth surface may lead to particle bounce-off. Natural surfaces, such as vegetation, whereas highly variable and often difficult to describe theoretically, generally promote dry deposition. 2

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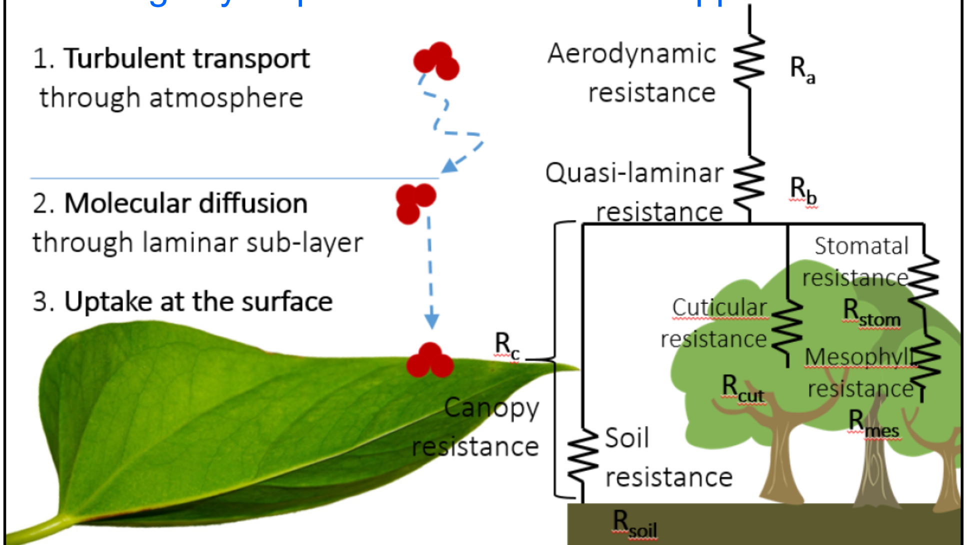
Modeling Dry Deposition: A Resistance Approach



Above a surface, a laminar layer will develop and fluid velocity will increase with distance from the surface, but not indefinitely. At some point, flow will become turbulent, with the laminar sublayer separating the turbulent layer from the surface 3 In the real world, most laminar boundary layers are extremely thin (order of 1 mm)

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Vertical Mixing Processes

The altitude of vertical mixing depends on atmospheric stability

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Dry deposition flux is directly proportional to the local concentration C of the depositing species, at some reference height above the surface (e.g., 10m or less)

$$F = -v_d C$$

The process of dry deposition of gases and particles is generally represented as consisting of three steps:

(1) aerodynamic transport down through the atmospheric surface layer to a very thin layer of stagnant air just adjacent to the surface;

(2) molecular (for gases) or Brownian (for particles) transport across this thin stagnant layer of air, called the *quasi-laminar sublayer,* to the surface itself;

(3) uptake at the surface.

Each of these steps contributes to the value of the deposition velocity v_d.

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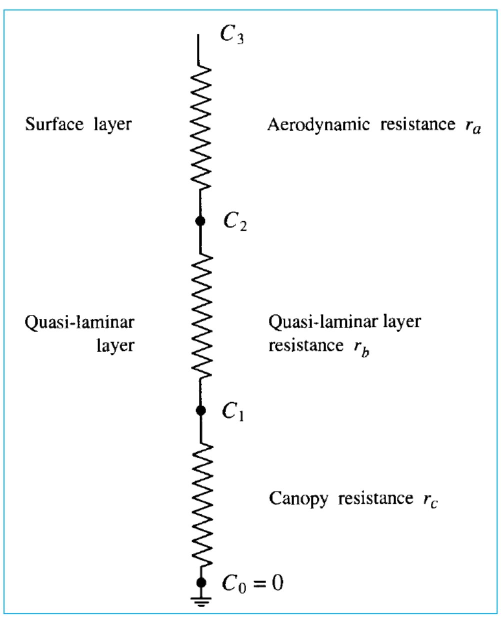
RESISTANCE MODEL FOR DRY DEPOSITION OF GASES



At steady state the overall flux of a vapor species is related to the concentration differences and resistances across the layers by

$$F = \frac{C_3 - C_2}{r_a} = \frac{C_2 - C_1}{r_b} = \frac{C_1 - C_0}{r_c} = \frac{C_3 - C_0}{r_t}$$

where $C_0 = 0$. Thus

$$\frac{C_3}{r_t} = \frac{C_3 - C_2}{r_a} = \frac{C_2 - C_1}{r_b} = \frac{C_1}{r_c}$$

By solving the three independent equations for the three unknowns, C1, C2, C3, one obtains

$$v_d^{-1} = r_t = r_a + r_b + r_c$$

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For **particles**, the model is identical to that for gases except that **particle settling operates in parallel with the three resistances in series**. It is usually assumed that particles adhere to the surface on contact so that the surface or canopy resistance r_c = 0. In this case the vertical flux is

$$F = \frac{C_3 - C_2}{r_a} + v_s C_3 = \frac{C_2 - C_1}{r_b} + v_s C_2 = \frac{C_3 - C_1}{r_t}$$

where, because of assumed perfect removal, $C_1 = 0$. Thus

$$\frac{C_3}{r_t} = \frac{C_3 - C_2}{r_a} + v_s C_3 = \frac{C_2}{r_b} + v_s C_2$$

$$v_d = \frac{1}{r_t} = \frac{1}{r_a + r_b + r_a r_b v_s} + v_s$$

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AERODYNAMIC RESISTANCE

Turbulent transport is the mechanism that brings material from the bulk atmosphere down to the surface and therefore determines the aerodynamic resistance. The turbulence intensity is dependent principally on the lower atmospheric stability and the surface roughness and can be determined from micrometeorological measurements and surface characteristics such as wind speed, temperature, and radiation and the surface roughness length.

Boundary layer meteorology

$$r_a = \begin{cases} \frac{1}{\kappa u_*}\left[\ln\left(\frac{z}{z_0}\right) + 4.7(\zeta - \zeta_0)\right] & \text{(stable)} \\ \frac{1}{\kappa u_*}\ln\left(\frac{z}{z_0}\right) & \text{(neutral)} \\ \frac{1}{\kappa u_*}\left[\ln\left(\frac{z}{z_0}\right) + \ln\left(\frac{(\eta_0^2+1)(\eta_0+1)^2}{(\eta_r^2+1)(\eta_r+1)^2}\right) + 2(\tan^{-1}\eta_r - \tan^{-1}\eta_0)\right] & \text{(unstable)} \end{cases}$$

where $\eta_0 = (1 - 15\zeta_0)^{1/4}$, $\eta_r = (1 - 15\zeta_r)^{1/4}$, $\zeta_0 = z_0/L$.

L is the Monin-Obukhov length; z_o *is* the roughness length; u_* is the friction velocity.[8]

QUASI-LAMINAR RESISTANCE

An expression for the overall quasi-laminar resistance for particles has been developed by Zhang et al. (2001)

$$r_b = \frac{1}{\varepsilon_0 u_* (E_B - E_{IM} + E_{IN}) R_1}$$

where

E_B = collection efficiency from Browian diffusion
E_{IM} = collection efficiency from impaction
E_{IN} = collection efficiency from interception

$$r_b = \frac{1}{3u_* \left[Sc^{-\gamma} + \left(\frac{St}{\alpha + St} \right)^2 + \frac{1}{2} \left(\frac{D_p}{A} \right)^2 \right] R_1}$$

where *R1* is a correction factor representing the fraction of particles that stick to the surface, St is the Stokes number, Sc is the Schmidt number. 9

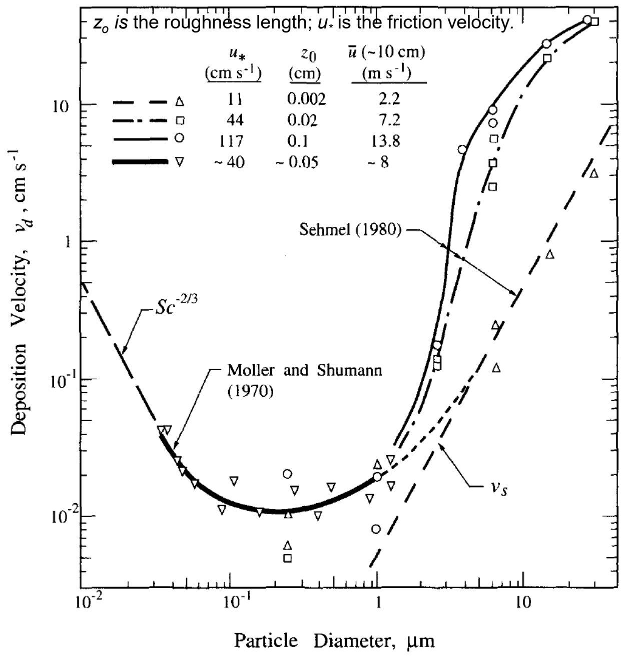
The aerodynamic and quasi-laminar resistances are affected by wind speed, vegetation height, leaf size, and atmospheric stability. Smaller resistances and hence higher deposition rates are expected over tall forests than over short grass. Also, smaller resistances are expected under unstable than under stable and neutral conditions. Typical aerodynamic layer resistances for a 4 m s-1 wind speed are as follows:

$z_0 = 0.1$ m	Grass	$r_a \simeq 60\ s\ m^{-1}$
$= 0.1$ m	Crop	$\simeq 20\ s\ m^{-1}$
$= 10$ m	Conifer forest	$\simeq 10\ s\ m^{-1}$

TABLE 19.1 Typical Dry Deposition Velocities for Some Atmospheric Gases

	v_d(cm s^{-1}) over		
Species	Continent	Ocean	Ice/Snow
CO	0.03	0	0
N_2O	0	0	0
NO	0.016	0.003	0.002
NO_2	0.1	0.02	0.01
HNO_3	4	1	0.5
O_3	0.4	0.07	0.07
H_2O_2	0.5	1	0.32

Source: Hauglustaine et al. (1994). 10



$$v_d = \frac{1}{r_t} = \frac{1}{r_a + r_b + r_a r_b v_s} + v_s$$

FIGURE 19.2 Particle dry deposition velocity data for deposition on a water surface in a wind tunnel (Slinn et al. 1978).

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For a spherical particle of density ρ_p in a fluid of density ρ, $m_p = (\pi/6)D_p^3(\rho_p - \rho)$, where the factor $(\rho_p - \rho)$ is needed to account for both gravity and buoyancy. However, since generally $\rho_p \gg \rho, m_p = (\pi/6)D_p^3\rho_p$ and (9.41) can be rewritten in the more convenient form:

$$v_t = \frac{1}{18}\frac{D_p^2 \rho_p g C_c}{\mu} \qquad (9.42)$$

$$C_c = 1 + \frac{2\lambda}{D_p}\left[1.257 + 0.4\exp\left(-\frac{1.1D_p}{2\lambda}\right)\right]$$

The viscosity of air (μ) depends mostly on the temperature. At 15 °C, the viscosity of air is 1.81 × 10^{-5} kg/(m·s) , 18.1 μPa·s or 1.81 × 10^{-5} Pa·s .

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Emerging research topics related to aerosol physics

https://www.forbes.com/sites/jamesconca/2019/09/10/solar-geoengineering-we-better-do-it-or-well-burn/?sh=1f5602b18add
Forbes: Why Solar Geoengineering May Be Our Only Hope To Reverse Global Warming

White, Science, 2025

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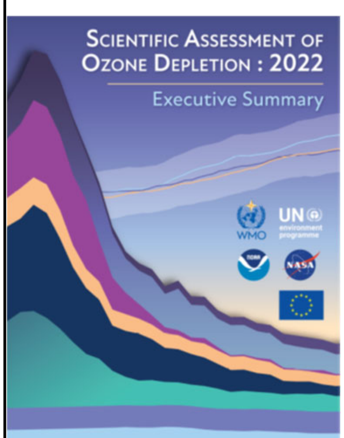
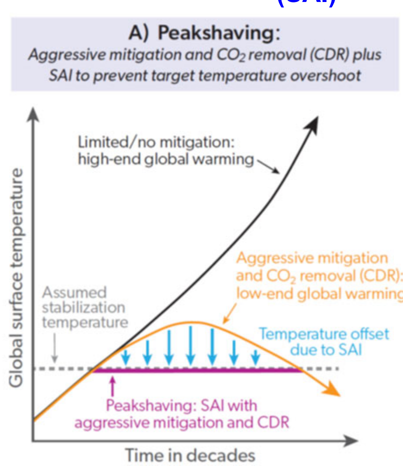
WMO 2022 Ozone Assessment Report has a specific chapter on **Stratospheric Aerosol Injection (SAI)**



Chapter 6:
Stratospheric Aerosol Injection and Its Potential Effect on the Stratospheric Ozone Layer



"**SAI** and other solar radiation modification (SRM) approaches may therefore **be the only option** to keep the global surface temperature **below the limit of 1.5 °C**."

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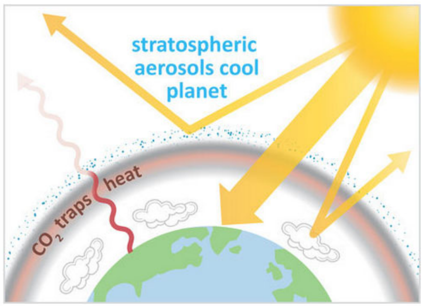
Geoengineering | Harvard's Sola

https://geoengineering.environment.harvard.edu/geoengineering

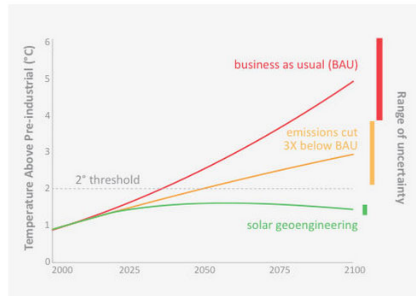
HARVARD UNIVERSITY

STAY IN TOUCH

HARVARD'S SOLAR GEOENGINEERING RESEARCH PROGRAM



SOLAR GEOENGINEERING



SOLAR GEOENGINEERING BENEFITS AND RISKS

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Solar geoengineering (solar radiation modification) via stratospheric aerosol injection is a good example of the application of <u>aerosol physics</u>

Question 1: How good do we understand the processes controlling particle evolution in the stratosphere? Any unknown unknowns?

Question 2: To be cost effective (to achieve the needed steady state AOD with minimum mass of aerosols to be injected), what is the ideal size of particles we shall inject? Does it depend on altitudes?

Question 3: How to do it?

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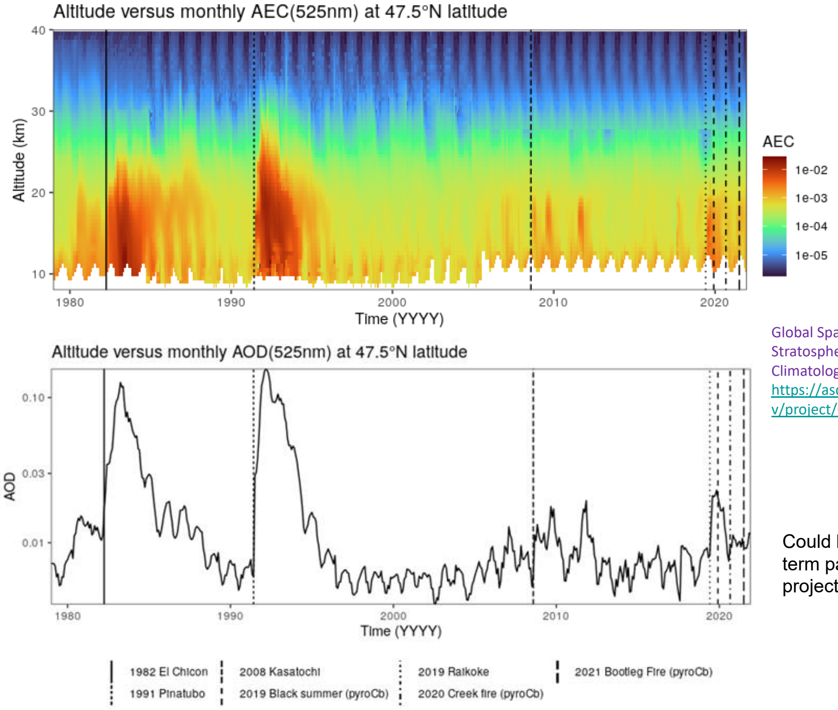

Global Space-based Stratospheric Aerosol Climatology (GloSSAC) https://asdc.larc.nasa.gov/project/GloSSAC

Could be a good term paper project

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Quaglia, I., et al.:Interactive Stratospheric Aerosol models response to different amount and altitude of SO_2 injections during the 1991 Pinatubo eruption, Atmos. Chem. Phys. Discuss. [preprint], https://doi.org/10.5194/acp-2022-514, in review, 2022.

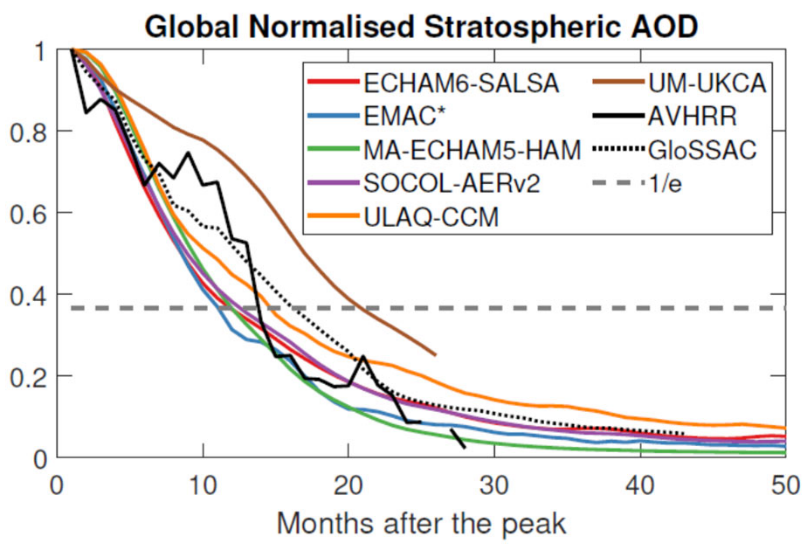

Figure 4. Time evolution of monthly values of the normalised global stratospheric AOD for models (colored lines) and AVHRR and GloSSAC observations (black lines). The dashed gray line represents the 1/e value. The experiment shown is Med-19km for ECHAM6-SALSA, ECHAM5-HAM, SOCOL-AERv2 and ULAQ-CCM, and Low-22km for UM-UKCA. For EMAC, it refers to the only experiment provided.

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Discussion:

What are the possible reasons of the faster decay or decrease of stratospheric aerosols after volcano eruptions predicted by various models?

What is the implication to the proposed stratospheric aerosol injection?

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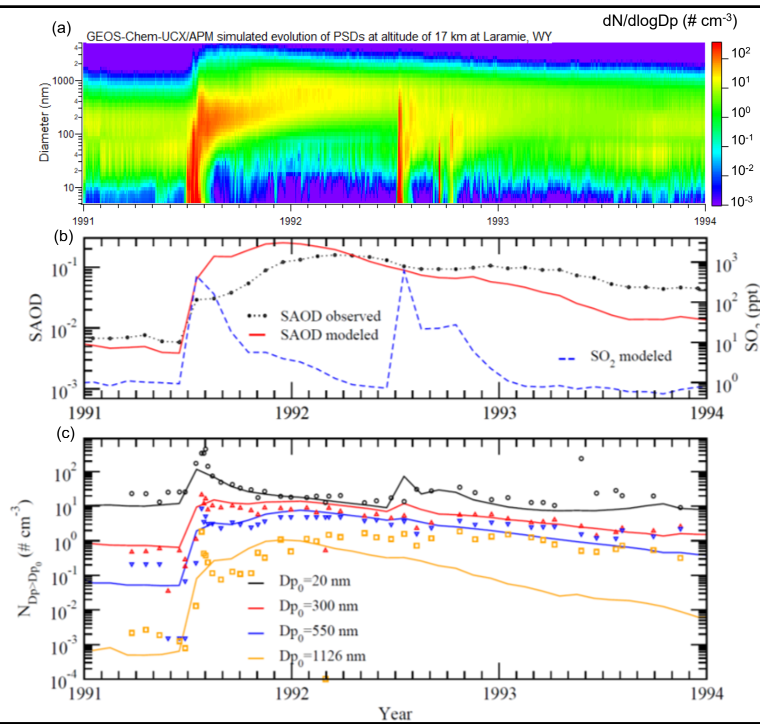

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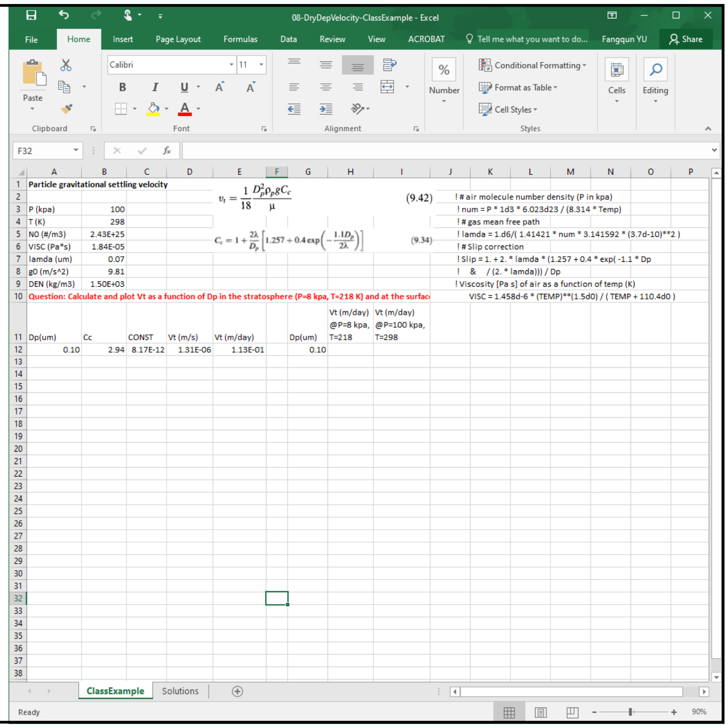
Particle gravitational settling velocity

Effect of altitudes (P, T) and particle sizes?

Stratosphere
P=8 kpa,
T=218 K

Surface
P=100 kpa
T=298 K



08-DryDepVelocity-ClassExample - Excel

Particle gravitational settling velocity

$$v_t = \frac{1}{18}\frac{D_p^2 \rho_p g C_c}{\mu} \quad (9.42)$$

$$C_c = 1 + \frac{2\lambda}{D_p}\left[1.257 + 0.4\exp\left(-\frac{1.1D_p}{2\lambda}\right)\right] \quad (9.34)$$

P (kpa)	100
T (K)	298
NO (#/m3)	2.43E+25
VISC (Pa*s)	1.84E-05
lamda (um)	0.07
g0 (m/s^2)	9.81
DEN (kg/m3)	1.50E+03

! # air molecule number density (P in kpa)
! num = P * 1d3 * 6.023d23 / (8.314 * Temp)
! # gas mean free path
! lamda = 1.d6/(1.41421 * num * 3.141592 * (3.7d-10)**2)
! # Slip correction
! Slip = 1. + 2. * lamda * (1.257 + 0.4 * exp(-1.1 * Dp
! & / (2. * lamda))) / Dp
! Viscosity [Pa s] of air as a function of temp (K)
VISC = 1.458d-6 * (TEMP)**(1.5d0) / (TEMP + 110.4d0)

Question: Calculate and plot Vt as a function of Dp in the stratosphere (P=8 kpa, T=218 K) and at the surfac

Dp(um)	Cc	CONST	Vt (m/s)	Vt (m/day)	Dp(um)	Vt (m/day) @P=8 kpa, T=218	Vt (m/day) @P=100 kpa, T=298
0.10	2.94	8.17E-12	1.31E-06	1.13E-01	0.10		

ClassExample | Solutions

Particle gravitational settling velocity

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Stratosphere
P=8 kpa,
T=218 K

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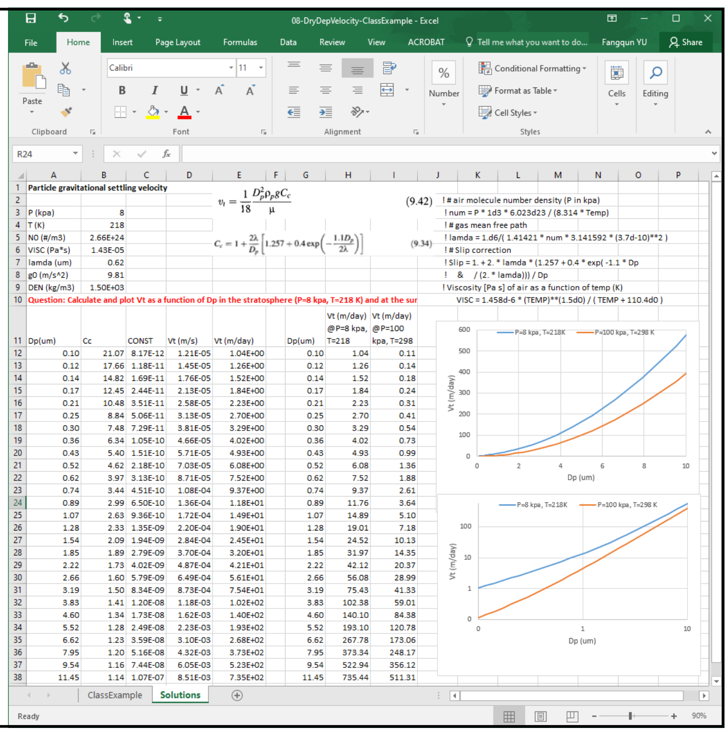
P=100 kpa
T=298 K



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Particle gravitational settling velocity

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$$C_c = 1 + \frac{2\lambda}{D_p}\left[1.257 + 0.4\exp\left(-\frac{1.1D_p}{2\lambda}\right)\right] \quad (9.34)$$

P (kpa)	8
T (K)	218
NO (#/m3)	2.66E+24
VISC (Pa*s)	1.43E-05
lamda (um)	0.62
g0 (m/s^2)	9.81
DEN (kg/m3)	1.50E+03

! # air molecule number density (P in kpa)
! num = P * 1d3 * 6.023d23 / (8.314 * Temp)
! # gas mean free path
! lamda = 1.d6/(1.41421 * num * 3.141592 * (3.7d-10)**2)
! # Slip correction
! Slip = 1. + 2. * lamda * (1.257 + 0.4 * exp(-1.1 * Dp
! & / (2. * lamda))) / Dp
! Viscosity [Pa s] of air as a function of temp (K)
VISC = 1.458d-6 * (TEMP)**(1.5d0) / (TEMP + 110.4d0)

Question: Calculate and plot Vt as a function of Dp in the stratosphere (P=8 kpa, T=218 K) and at the sur

Dp(um)	Cc	CONST	Vt (m/s)	Vt (m/day)	Dp(um)	Vt (m/day) @P=8 kpa, T=218	Vt (m/day) @P=100 kpa, T=298
0.10	21.07	8.17E-12	1.21E-05	1.04E+00	0.10	1.04	0.11
0.12	17.66	1.18E-11	1.45E-05	1.26E+00	0.12	1.26	0.14
0.14	14.82	1.69E-11	1.76E-05	1.52E+00	0.14	1.52	0.18
0.17	12.45	2.44E-11	2.13E-05	1.84E+00	0.17	1.84	0.24
0.21	10.48	3.51E-11	2.58E-05	2.23E+00	0.21	2.23	0.31
0.25	8.84	5.06E-11	3.13E-05	2.70E+00	0.25	2.70	0.41
0.30	7.48	7.29E-11	3.81E-05	3.29E+00	0.30	3.29	0.54
0.36	6.34	1.05E-10	4.66E-05	4.02E+00	0.36	4.02	0.73
0.43	5.40	1.51E-10	5.71E-05	4.93E+00	0.43	4.93	0.99
0.52	4.62	2.18E-10	7.03E-05	6.08E+00	0.52	6.08	1.36
0.62	3.97	3.13E-10	8.71E-05	7.52E+00	0.62	7.52	1.88
0.74	3.44	4.51E-10	1.08E-04	9.37E+00	0.74	9.37	2.61
0.89	2.99	6.50E-10	1.36E-04	1.18E+01	0.89	11.76	3.64
1.07	2.63	9.36E-10	1.72E-04	1.49E+01	1.07	14.89	5.10
1.28	2.33	1.35E-09	2.20E-04	1.90E+01	1.28	19.01	7.18
1.54	2.09	1.94E-09	2.84E-04	2.45E+01	1.54	24.52	10.13
1.85	1.89	2.79E-09	3.70E-04	3.20E+01	1.85	31.97	14.35
2.22	1.73	4.02E-09	4.87E-04	4.21E+01	2.22	42.12	20.37
2.66	1.60	5.79E-09	6.49E-04	5.61E+01	2.66	56.08	28.99
3.19	1.50	8.34E-09	8.73E-04	7.54E+01	3.19	75.43	41.33
3.83	1.41	1.20E-08	1.18E-03	1.02E+02	3.83	102.38	59.01
4.60	1.34	1.73E-08	1.62E-03	1.40E+02	4.60	140.10	84.38
5.52	1.28	2.49E-08	2.23E-03	1.93E+02	5.52	193.10	120.78
6.62	1.23	3.59E-08	3.10E-03	2.68E+02	6.62	267.78	173.06
7.95	1.20	5.16E-08	4.32E-03	3.73E+02	7.95	373.34	248.17
9.54	1.16	7.44E-08	6.05E-03	5.23E+02	9.54	522.94	356.12
11.45	1.14	1.07E-07	8.51E-03	7.35E+02	11.45	735.44	511.31

ClassExample | Solutions