

惑星大気GCMの物理過程

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自己紹介

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東京大学理学部地球惑星物理学科卒業

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地球惑星科学専攻 大学院生(修士→博士)
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(GCOE→リーディング大学院)
現在次のポスト探し中です...

主な研究業績

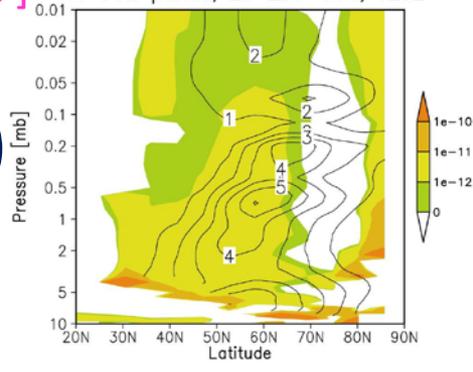
[Kuroda et al., 2007]

Refractive index & T amplitude (s=1)
6.6 Sols period, Ls=280-300, TES2

火星GCMを用いた研究(主に大気力学)

北極域傾圧不安定波の鉛直伝播

[Kuroda et al., 2007, GRL 34, L09203]

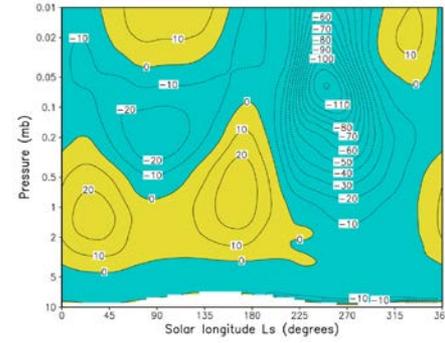


赤道半年振動[Kuroda et al., 2008, GRL 35, L23202]

[Kuroda et al., 2008]

全球ダストストームによる極夜の昇温

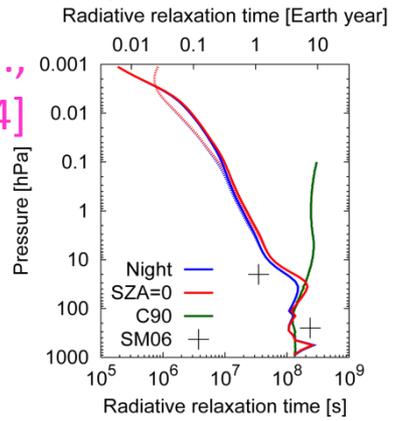
[Kuroda et al., 2009, JMSJ 87, 913-921]



北極極夜のCO₂降雪[Kuroda et al., 2013,

GRL 40, 1484-1488]

[Kuroda et al., 2014]

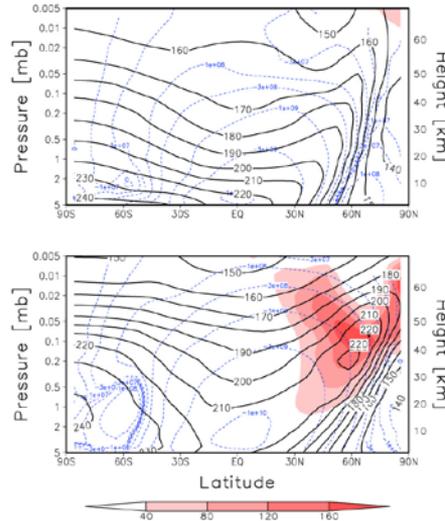


木星大気にも進出中

成層圏放射コードの開発と
放射緩和時間

[Kuroda et al., 2014, Icarus 242, 149-157]

[Kuroda et al., 2009]



金星GCMにも着手(東北大の学生と)

火星ダスト気象についてのレビュー論文(2011年)

Aeolian Research 3 (2011) 145–156



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Review Article

Influence of dust on the dynamics of the martian atmosphere above the first scale height

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ABSTRACT

Dust suspended in the martian atmosphere strongly affects the radiative transfer. Diabatic heating and cooling it creates are prominent factors that drive the atmosphere at various scales. This paper provides a review of dust influence on the large-scale dynamics in the atmosphere of Mars above approximately 10 km. We outline the established properties of dust that influence the diabatic heating/cooling rates, and summarize the current knowledge of dust-related effects on the zonal-mean circulation and zonally asymmetric disturbances: planetary waves and tides.

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これまでの開発履歴

火星GCM(大学院生時代～)

力学コア

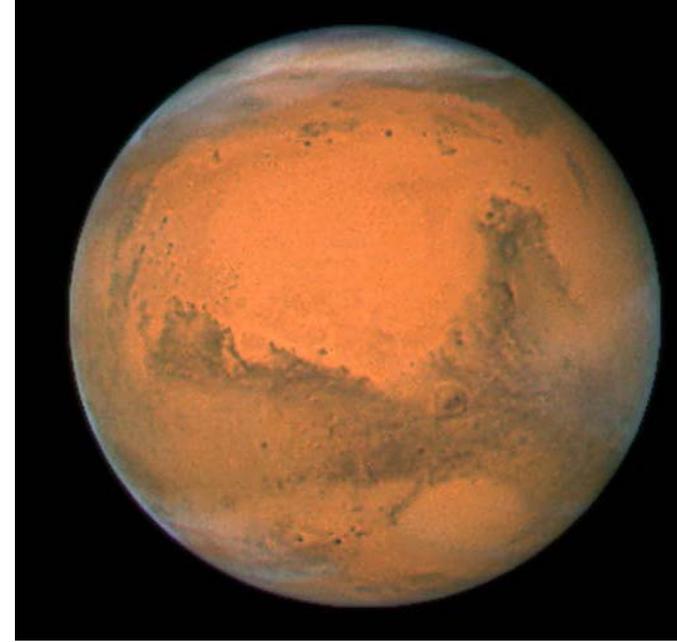
CCSR/NIES(/FRCGC) AGCM

5.4.02→5.4g(D論)→5.6→5.7b (MIROC)

3次元プリミティブ方程式(スペクトル法)

2010年頃より「DRAMATIC MGCM」と呼び始める

(DRAMATIC = Dynamics, RAdiation, MAterial Transport and their mutual InteraCtions)



分解能

水平 : T21($\sim 5.6^\circ \times 5.6^\circ$, 赤道域グリッド間隔 $\sim 333\text{km}$)

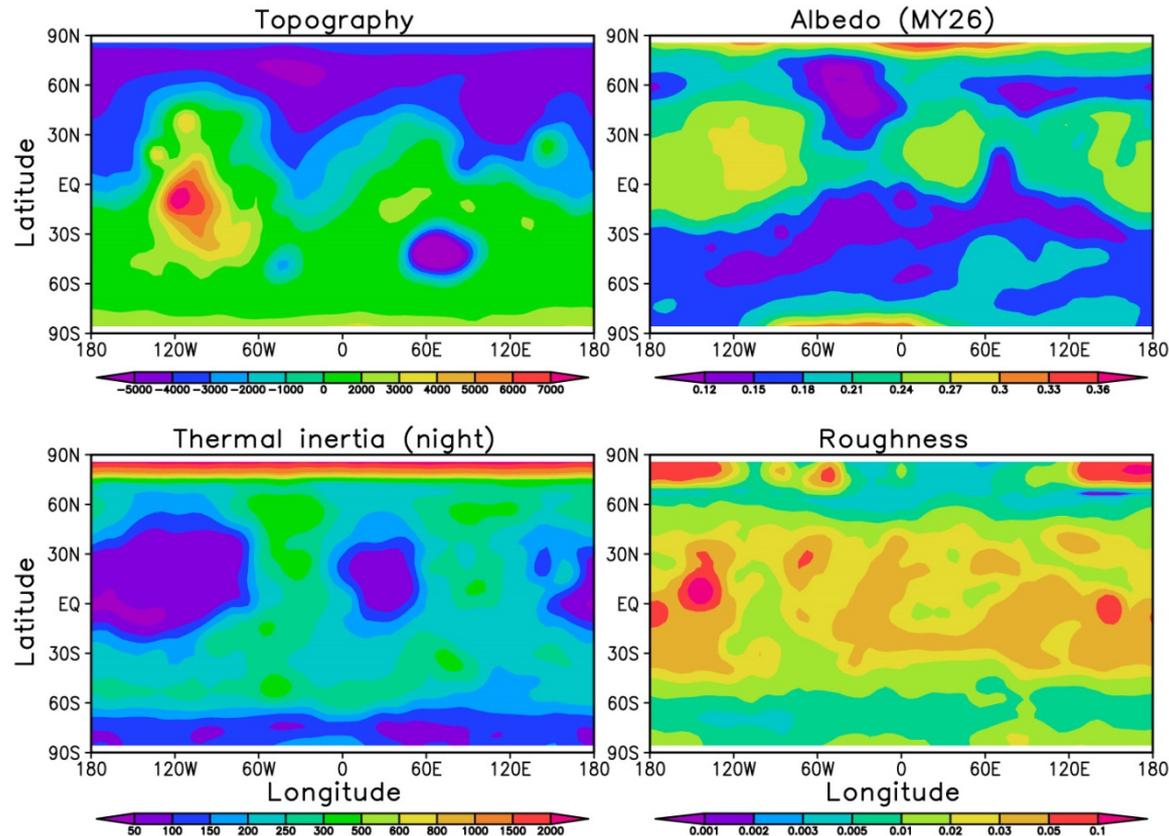
鉛直 : σ レベル30層 \rightarrow 69層, モデル上端 $\sim 80\text{km} \rightarrow \sim 100\text{km}$

地表面パラメータ

地形: MGS-MOLA観測 [Smith et al., 1999]

アルベド・熱慣性: Mars Climate Database v3 [Forget et al., 1999, 2001]→MGS-TES観測 [Putzig and Mellon, 2007]

粗度: MGS-MOLAデータより導出 [Heavens et al., 2008]
(→MGS-TESデータ [Hebrard et al., 2012])



CO₂相変化(1)

凝結したドライアイスは重力沈降により落下する
熱力学的効果(潜熱の放出、落下によるエネルギー交換)と
質量の交換(大気⇔ドライアイス)を導入

If T^* (predicted temperature) $< T_c$ (condensation temperature)

$$T^* = T_c$$

$$\delta m_N = \frac{c_p M_N}{L} (T_{c_N} - T_N^*)$$

$$\delta m_l = \frac{c_p M_l}{L} (T_{c_l} - T_l^*) - \frac{1}{L} \left[\underbrace{g(z_{l+1} - z_l)}_{\textcircled{2}} + \underbrace{c_{ice} (T_{c_{l+1}} - T_{c_l})}_{\textcircled{3}} \right] \sum_{k=l+1}^N \delta m_k \quad (\text{in } l \leq N-1)$$

If all the falling ice sublimates in layer l (i.e. $-\delta m_l > \sum_{k=l+1}^N \delta m_k$),

$$T_l = T_l^* + \frac{1}{c_p M_l} \left[-L + g(z_{l+1} - z_l) + c_{ice} (T_{c_{l+1}} - T_{c_l}) \right] \sum_{k=l+1}^N \delta m_k$$

At surface ($l=0$), $c_p M_l$ is replaced by $c_s A$,

and $\sum_{k=0}^N \delta m_k$ is added to the ground ice amount m_0 .

If the ground ice completely sublimates (i.e. $-\delta m_0 > m_0 + \sum_{k=1}^N \delta m_k$),

$$T_0 = T_0^* - \frac{L}{c_s A} m_0 - \frac{1}{c_s A} \left[L - g(z_{l+1} - z_l) - c_{ice} (T_{c_1} - T_{c_0}) \right] \sum_{k=1}^N \delta m_k$$

Accounting for the loss of atmospheric mass due to condensation,

$$\delta p_0 = -\frac{g}{A} \sum_{k=0}^N \delta m_k$$

CO₂総量(大気+ドライアイス雲・極冠)
の保存のため、地表面気圧を修正

① 潜熱の放出

② 沈降に伴うポテンシャルエネルギー
の放出

③ ドライアイスを下の層の凝結温度まで
温めるのに要するエネルギー

当初は凝結したドライアイスは地面まで即時に沈降させる仕様だった[Forget et al., 1998]

→結果は計算時間ステップに大きく依存し、水平・鉛直分解能や上端高度を変更すると全く計算結果が変わる

CO₂相変化(2)

重力沈降速度を導入(2010年頃)

変形ストークス沈降[Kasten, 1968]

$$v_{sed} = \frac{2\rho g r^2}{9\eta} \left[1 + \frac{\lambda}{r} \left\{ A + B \exp\left(-\frac{Cr}{\lambda}\right) \right\} \right]$$

η : 大気粘性(温度依存)

λ : 平均自由行程(温度・密度依存)

r : 粒子半径

ρ : 大気密度

A=1.15, B=0.497, C=0.92 for CO₂ air

[Rader, 1990]

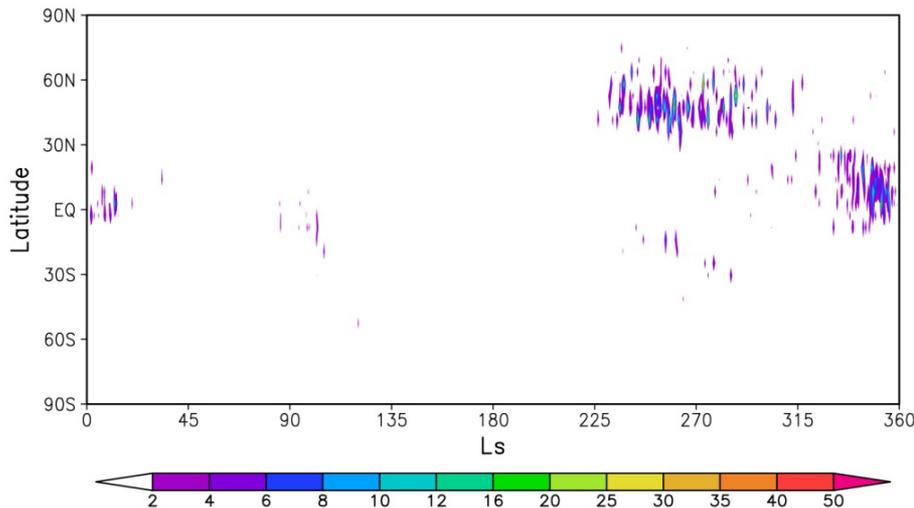
粒径の定義

r_0 : 0m高度での粒径(=50 μ m, 参考: Hayne et al. [2012]など)

h : 粒径スケールハイト(=20km)

$$r(z) = r_0 \exp(-z/h)$$

→高度約78kmで粒径1 μ mとなり、中層CO₂氷雲の観測と整合(参考: Maattanen et al. [2010]など)



1Pa(~64km高度)でのCO₂氷雲分布(ppm of mass)

北半球春～夏の赤道域、北半球冬の中緯度域に雲生成が見られる: 観測と整合

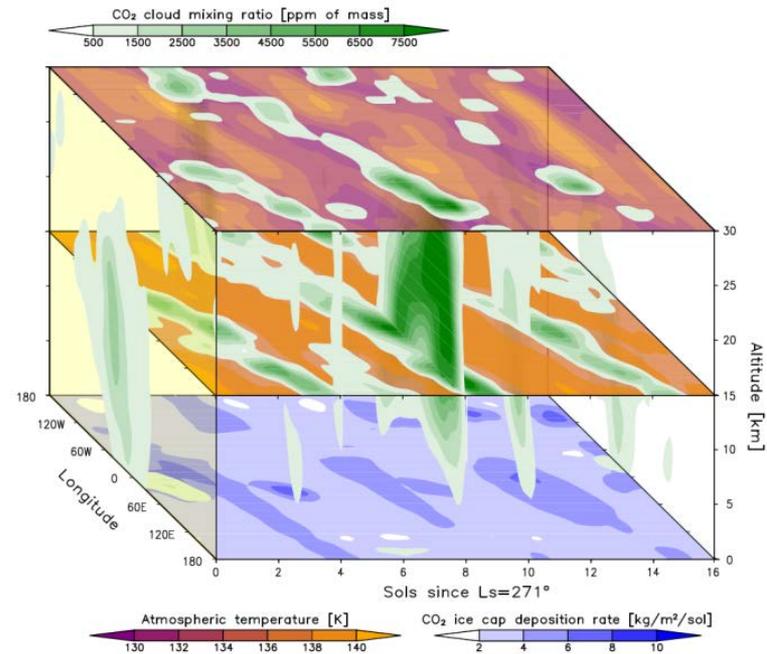
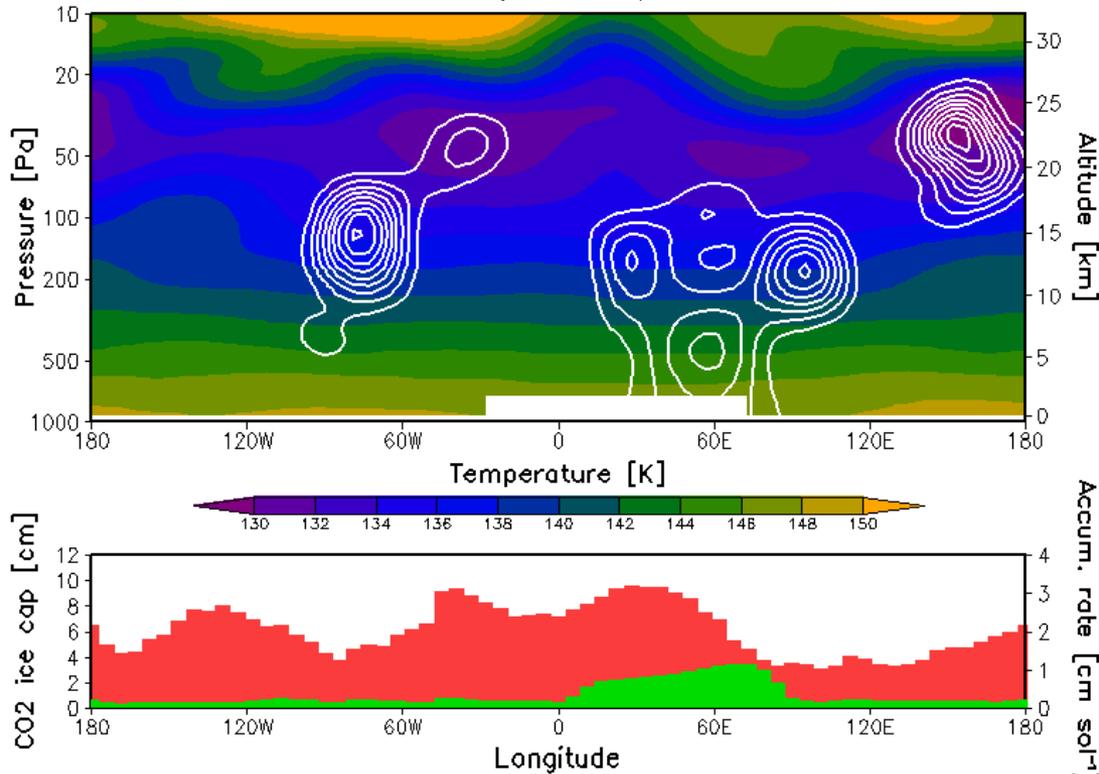
[Maattanen et al., 2010;

McConnochie et al., 2010]

冬の北極域のCO₂降雪[Kuroda et al., 2013]

(朝日新聞 2013年5月13日)

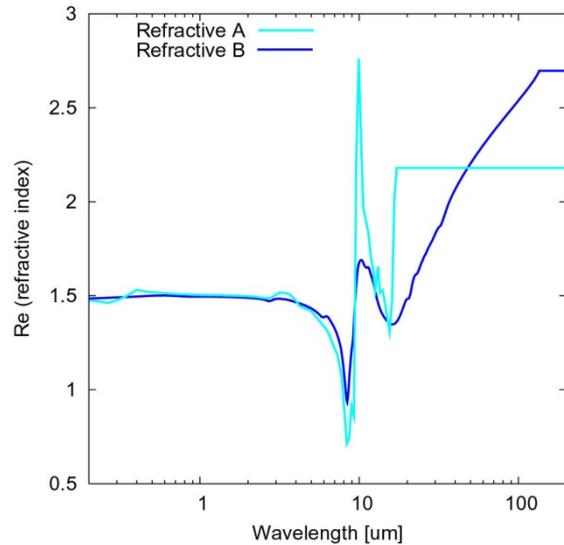
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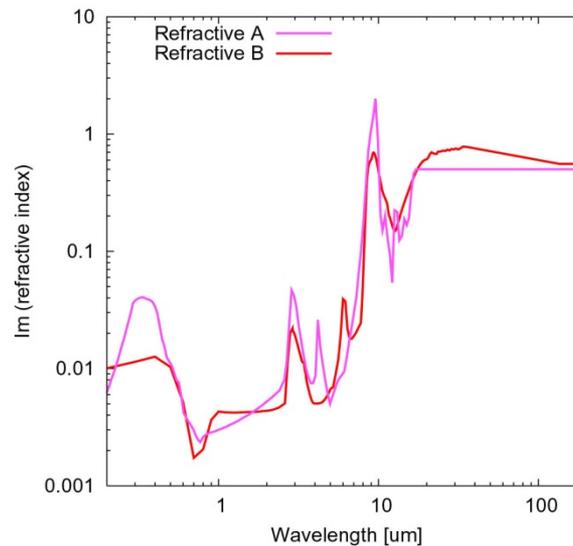
ダスト放射

複素屈折率と粒径について火星での観測に基づいたものを仮定して、mstrnXに組み込んでいる

実部



虚部



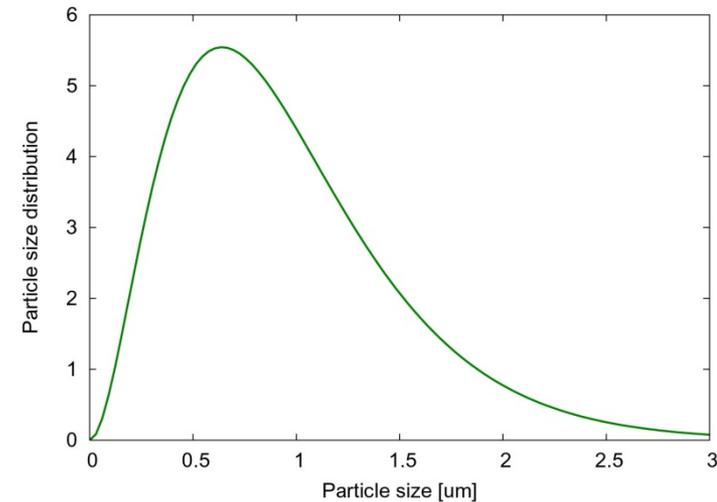
複素屈折率:

Wolff and Clancy [2003]より
(Refractive B)

Refractive Aはそれまでの
火星ダスト標準パラメータ
[Ockert-Bell et al., 1997;
Toon et al., 1977; Forget,
1998]

粒径分布: Tomasko et al. [1999]より
($r_{\text{eff}}=1.6 \mu\text{m}$, $v_{\text{eff}}=0.2 \mu\text{m}$)

(変形ガンマ関数) $n(r) = cr^{(1-3v_{\text{eff}})/v_{\text{eff}}} \exp[-r/r_{\text{eff}}v_{\text{eff}}]$

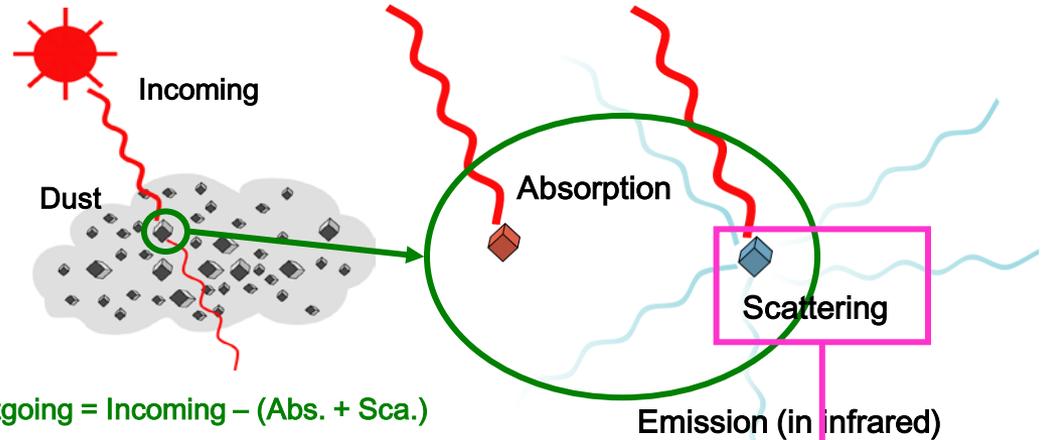


ダスト吸収・散乱係数 & 位相関数の計算(ミ-散乱の理論)

複素屈折率 $m = m_r - im_i$

$$a_n = \frac{u_n(x)u_n'(y) - mu_n(y)u_n'(x)}{w_n(x)u_n'(y) - mu_n(y)w_n'(x)}$$

$$b_n = \frac{mu_n(x)u_n'(y) - u_n(y)u_n'(x)}{mw_n(x)u_n'(y) - u_n(y)w_n'(x)}$$



Outgoing = Incoming - (Abs. + Sca.)

$$x = \frac{2\pi r}{\lambda} \quad y = mx = \frac{2\pi nr}{\lambda} \quad u_n(x) = \sqrt{\frac{\pi x}{2}} J_{n+\frac{1}{2}}(x) \quad w_n(x) = \sqrt{\frac{\pi x}{2}} H_{n+\frac{1}{2}}^{(2)}(x)$$

(ベッセル関数)

振幅関数

$$S_1(\theta) = \sum_{n=1}^{\infty} \frac{2n+1}{n(n+1)} \left[a_n P_n'(\cos \theta) + b_n \left\{ \cos \theta P_n'(\cos \theta) - \sin^2 \theta P_n''(\cos \theta) \right\} \right]$$

$$S_2(\theta) = \sum_{n=1}^{\infty} \frac{2n+1}{n(n+1)} \left[b_n P_n'(\cos \theta) + a_n \left\{ \cos \theta P_n'(\cos \theta) - \sin^2 \theta P_n''(\cos \theta) \right\} \right]$$

(ルジャンドル関数)

散乱断面積

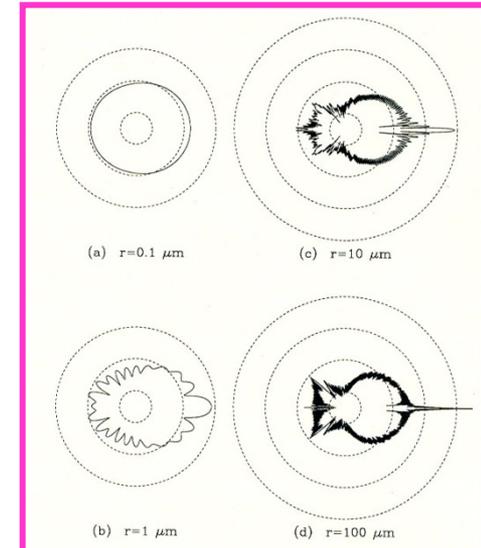
$$\sigma_s = \frac{\lambda^2}{4\pi} [S_1(\theta)S_1^*(\theta) + S_2(\theta)S_2^*(\theta)]$$

消散断面積

$$\sigma_e = \frac{\lambda^2}{4\pi} \text{Re}[S(0)]$$

ミ-散乱 粒径依存の例
(水滴、0.55 μm)

(「光の気象学」柴田清孝著、朝倉書店)



位相関数

$$P(\theta) = \frac{\lambda^2}{2\pi\sigma_s} [S_1(\theta)S_1^*(\theta) + S_2(\theta)S_2^*(\theta)]$$

散乱効率

$$Q_s = \frac{\sigma_s}{\pi r^2} = \frac{2}{x^2} \sum_{n=1}^{\infty} (2n+1)(a_n a_n^* + b_n b_n^*)$$

消散効率

$$Q_e = \frac{\sigma_e}{\pi r^2} = \frac{2}{x^2} \sum_{n=1}^{\infty} (2n+1) \operatorname{Re}[a_n + b_n]$$

吸収効率

$$Q_a = Q_e - Q_s$$

一次散乱アルベド

$$\omega_0 = \frac{Q_s}{Q_e} = \frac{\sigma_s}{\sigma_e}$$

以上は単一散乱についての計算。単位体積当たりの消散・吸収・散乱係数は、粒径分布を用いて以下の通りになる。

消散係数

粒径分布

吸収係数

散乱係数

$$\beta_e = \int_0^{\infty} \sigma_e(r) \frac{dn(r)}{dr} dr$$

$$\beta_a = \int_0^{\infty} \sigma_a(r) \frac{dn(r)}{dr} dr$$

$$\beta_s = \int_0^{\infty} \sigma_s(r) \frac{dn(r)}{dr} dr$$

位相関数

一次散乱アルベド

光学的厚さ

$$P(\theta) = \frac{1}{\beta_s} \int_0^{\infty} \sigma_s(r) P(r, \theta) \frac{dn(r)}{dr} dr$$

$$\omega_0 = \frac{\beta_s}{\beta_e}$$

$$\tau = \int_z^{\infty} \beta_e dz$$

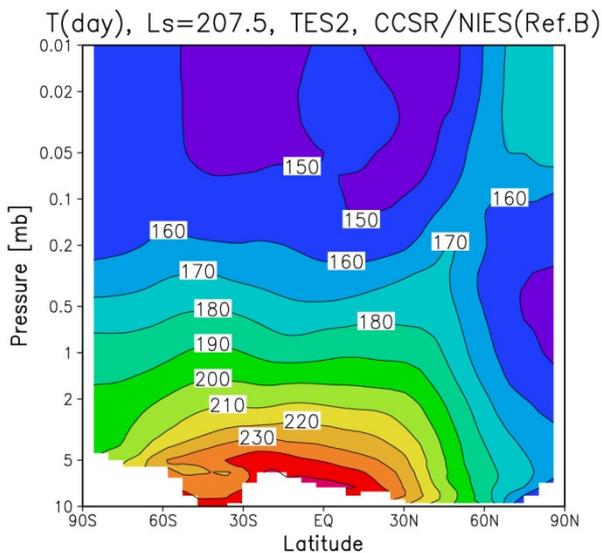
これらを用いて2ストリーム計算[Nakajima and Tanaka, 1986]

ダスト放射効果の評価

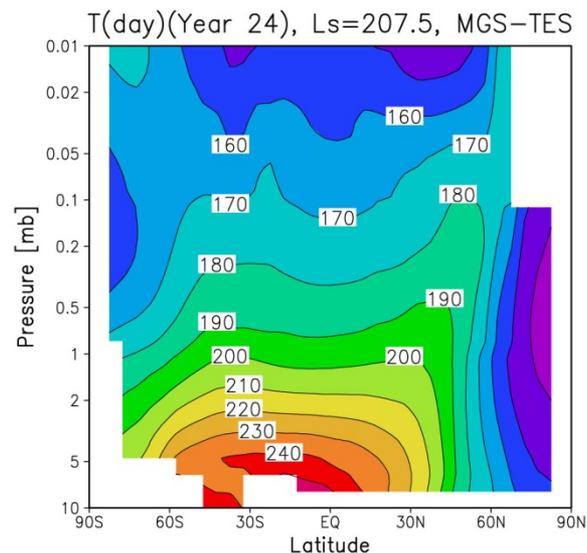
LS=207.5°
(北半球秋季)

$\tau_{\text{dust}} \sim 0.2$ (MY24相当)
daytime

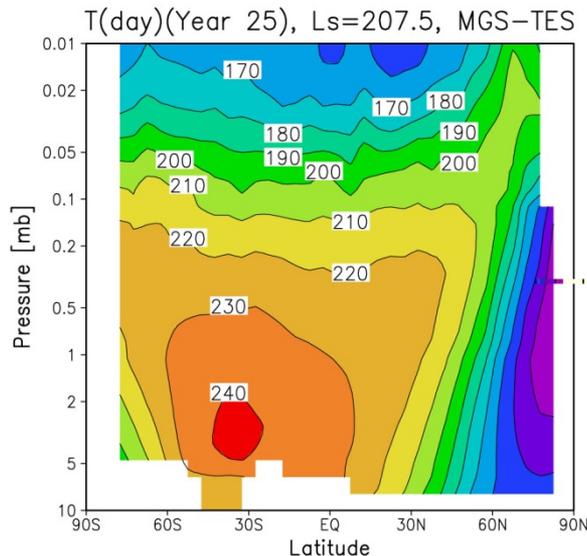
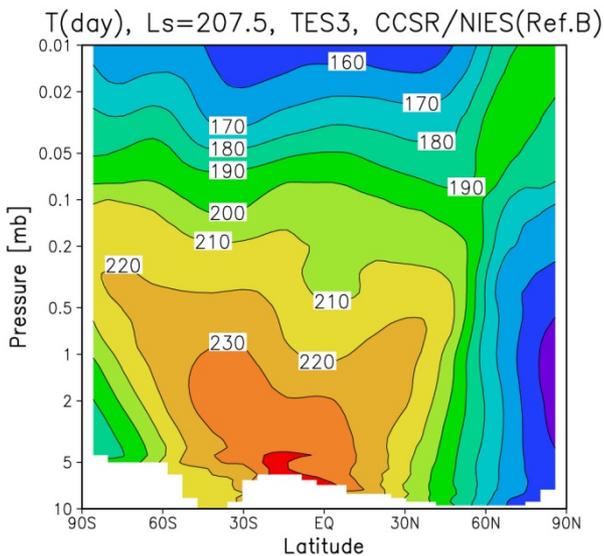
MGCM



MGS-TES観測

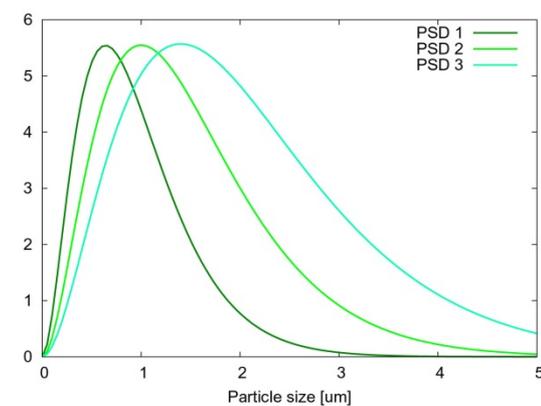
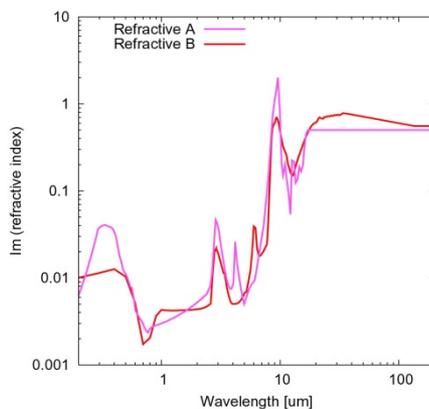


$\tau_{\text{dust}} \sim 2.2$ (MY25相当)
daytime



ダスト放射効果の感度実験 (異なる複素屈折率と粒径 分布での結果の違い)

$\tau_{\text{dust}} \sim 2.2$, $L_s = 207.5^\circ$, daytime



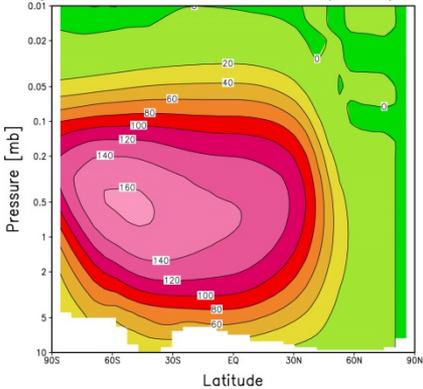
Refractive B,
PSD 1

Refractive A,
PSD 1

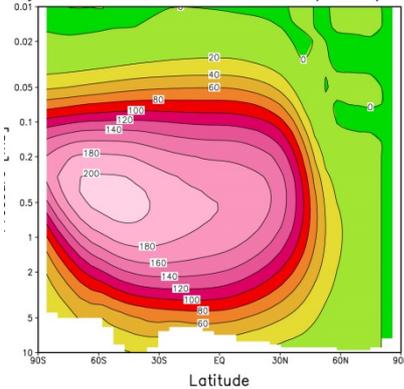
Refractive A,
PSD 2

Refractive A,
PSD 3

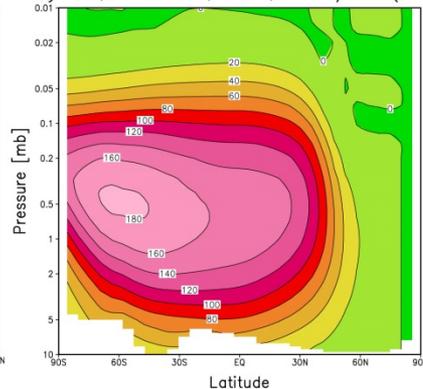
Dust shortwave heating rate
Daytime, $L_s = 207.5$, TES3, CCSR/NIES(Ref.B)



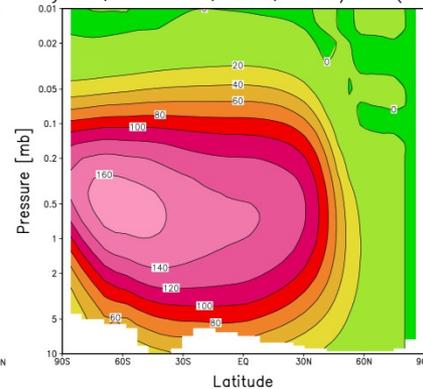
Dust shortwave heating rate
Daytime, $L_s = 207.5$, TES3, CCSR/NIES(Ref.A)



Dust shortwave heating rate
Daytime, $L_s = 207.5$, TES3, CCSR/NIES(Ref.A)

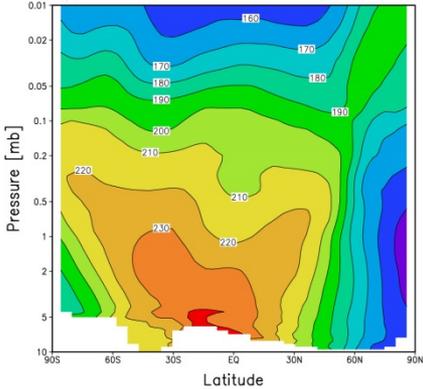


Dust shortwave heating rate
Daytime, $L_s = 207.5$, TES3, CCSR/NIES(Ref.A)

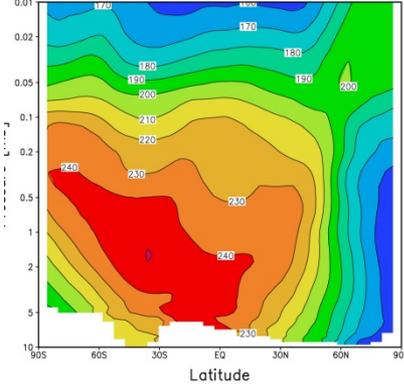


短波
放射
加熱
率

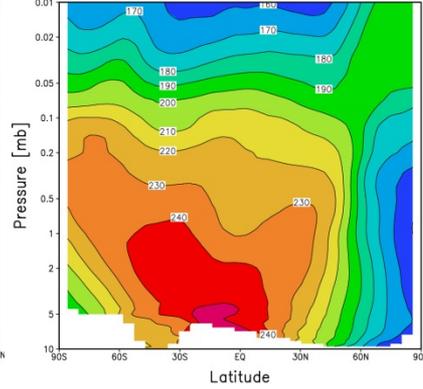
T(day), $L_s = 207.5$, TES3, CCSR/NIES(Ref.B)



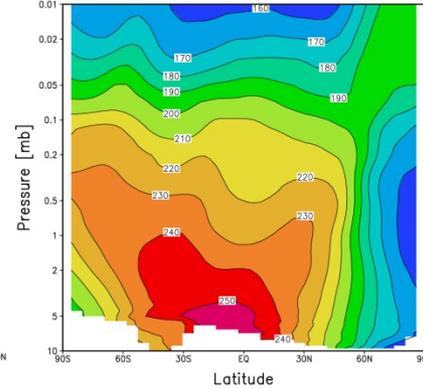
T(day), $L_s = 207.5$, TES3, CCSR/NIES(Ref.A)



T(day), $L_s = 207.5$, TES3, CCSR/NIES(Ref.A)



T(day), $L_s = 207.5$, TES3, CCSR/NIES(Ref.A)



温度

水循環(2008年～)

大規模凝結スキームの利用
(飽和水蒸気量を超えた水蒸気は凝結して氷雲になる)

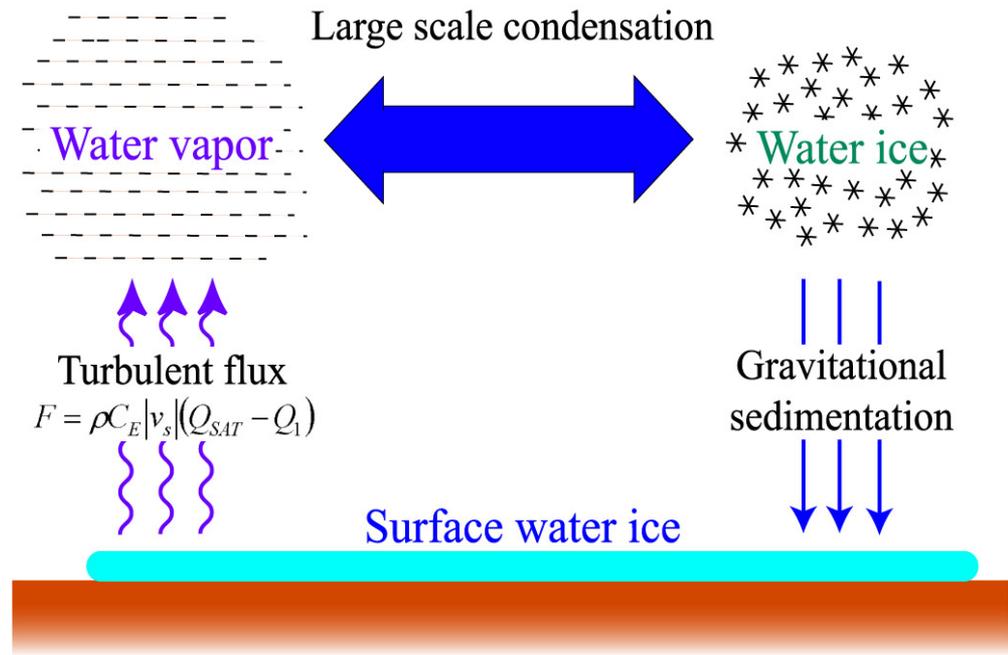
氷雲の重力沈降を導入
(CO₂氷雲と同じ変形ストークス沈降)

地表面に達した氷雲は地表氷として堆積

地表氷が存在する領域ではアルベドを0.3以上に設定

Turbulent fluxによる地表氷の昇華

(地表面レゴリスによる水の吸着、氷雲・水蒸気の放射効果は現在のところ未導入)



Montmessin et al. [2004]をもとにした
氷雲粒径の推定(過飽和を考慮せず)

$$r_c = \left(\frac{M_c}{(4/3)\pi\rho_i N} + r_0^3 \right)^{1/3}$$

r_c : 氷雲の半径
 M_c : 層内の氷の質量
 ρ_i : 氷の密度(917 kg m⁻³)
 N : 層内のダスト粒の数
 r_0 : ダスト粒の半径

$$N = \frac{M_d}{(4/3)\pi\rho_d r_0^3}$$

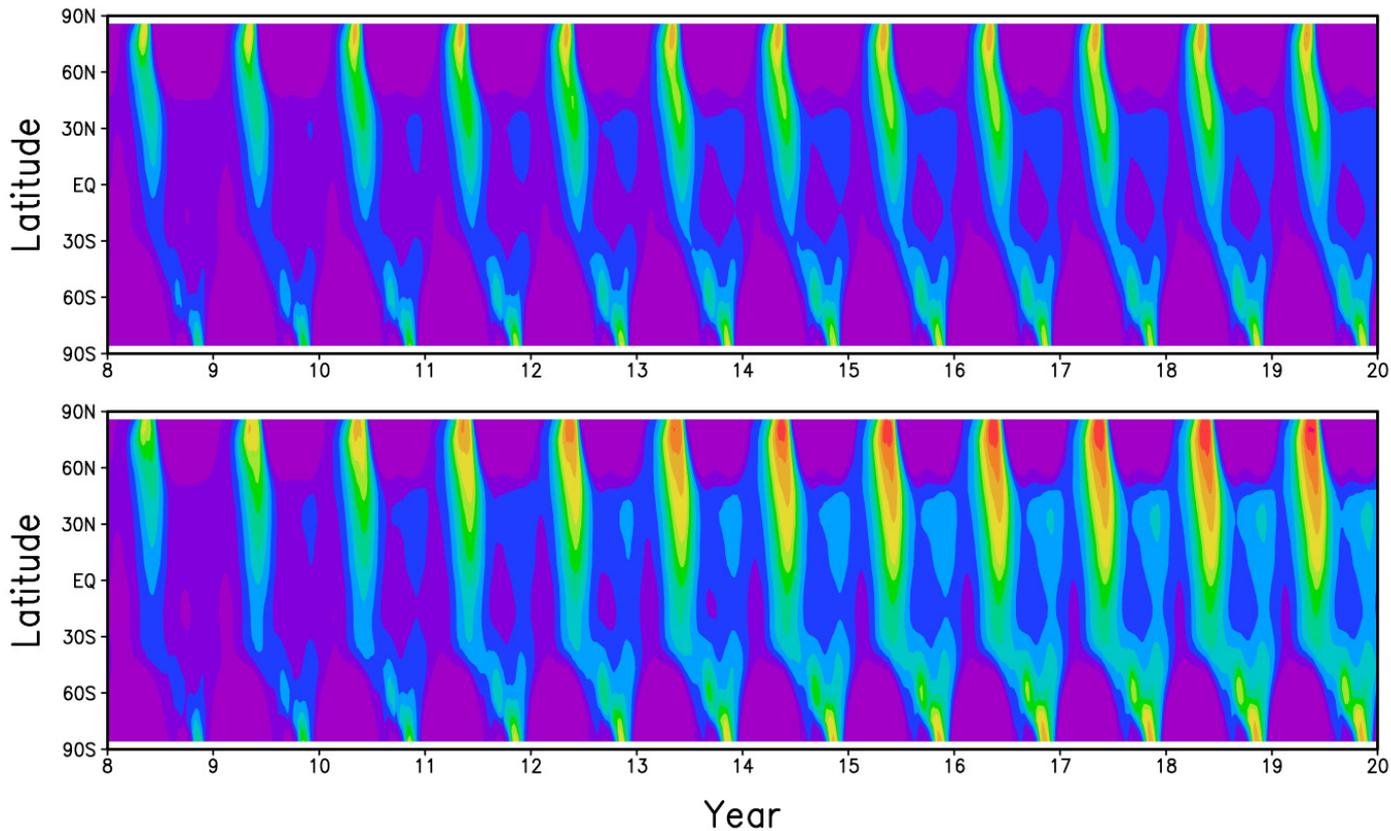
M_d : 層内のダストの質量
 ρ_d : ダストの密度(2500 kg m⁻³)

$$r_0(z) = r_{z0} \exp(-z/h)$$

r_{z0} : 高度0kmにおける半径(0.8μm)
 h : 粒径を定義するスケールハイト(18km)

水蒸気カラム量の結果(等温静止から8~20火星年)

過飽和なし



10倍過飽和を考慮(単に飽和水蒸気量を理論値の10倍に設定)

等温静止乾燥状態から計算スタート、北緯80度以北を氷床とし、そこから水が昇華して大気中に取り込まれてゆく設定

10倍過飽和を示唆する観測(MEx-SPICAM, Maltagliati et al. [2011])

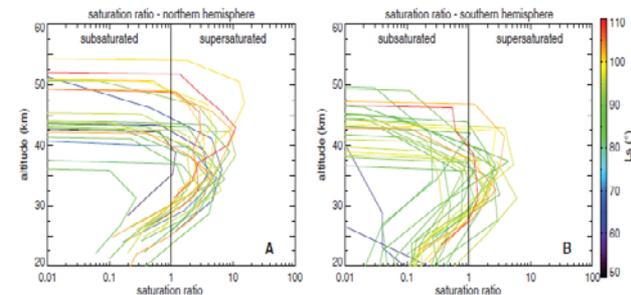
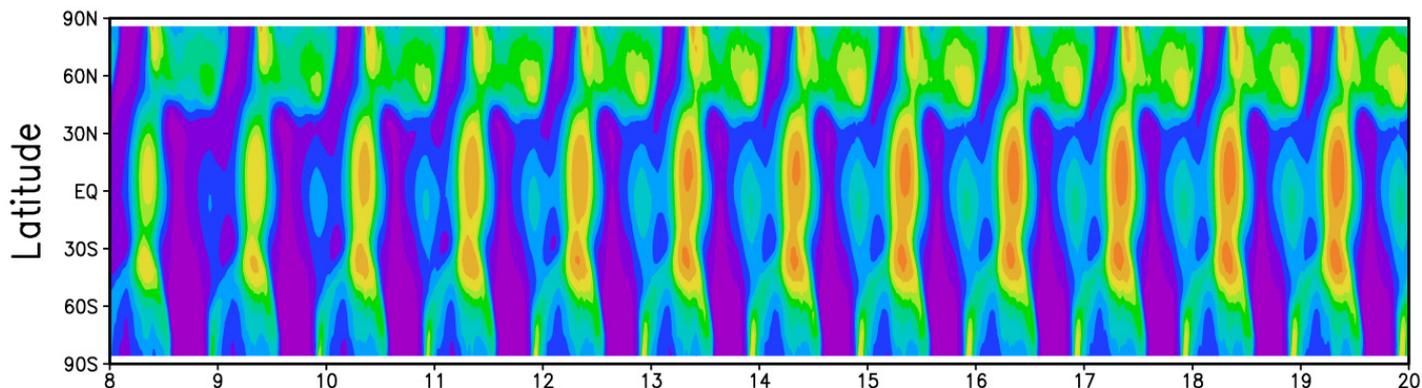


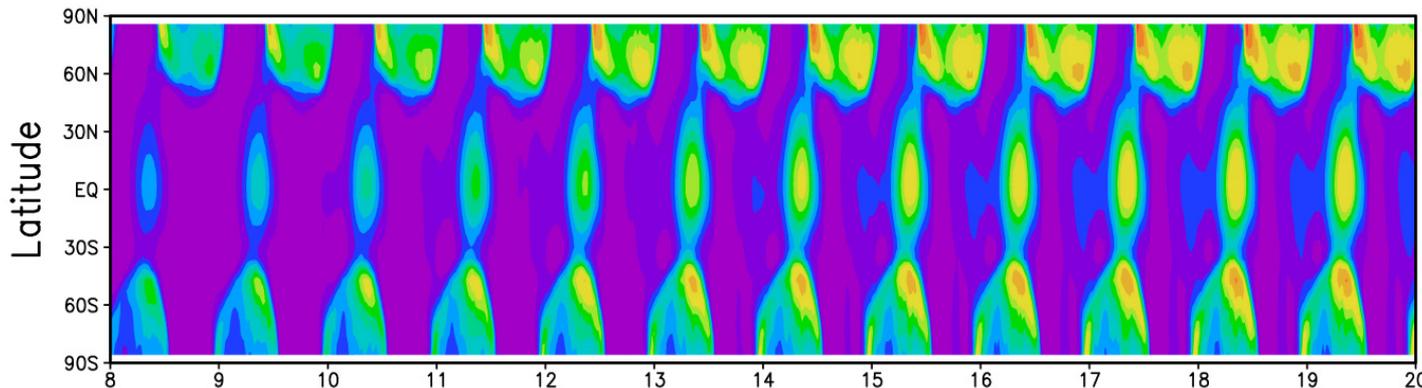
Fig. 3. Saturation ratio for all orbits of the campaign. (A) Northern hemisphere. (B) Southern hemisphere. The vertical line marks the value of 1, which corresponds to the saturated state.

氷雲光学的厚さの結果(等温静止から8~20火星年)

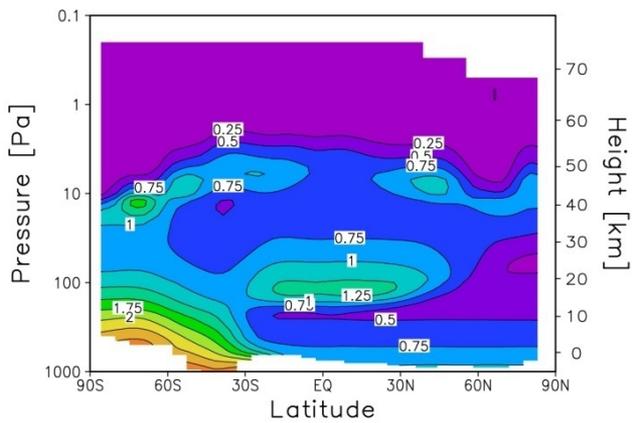
過飽和なし



10倍過飽和



LS=90°における
氷雲粒径(過飽和なし)



現状では10倍過飽和を考慮しても、
観測よりもかなり厚い(約20倍...)

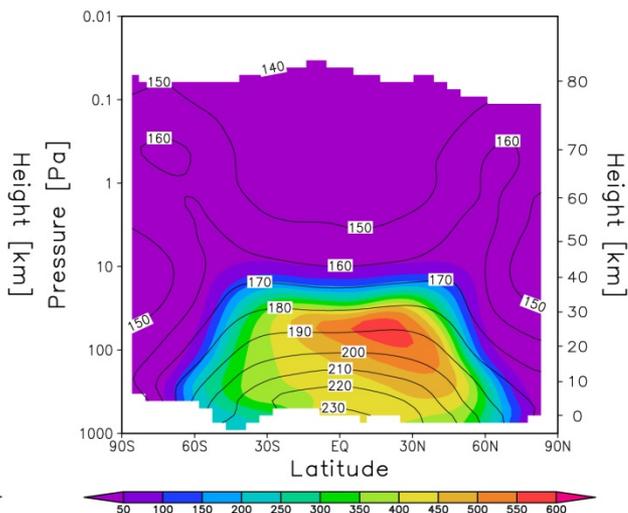
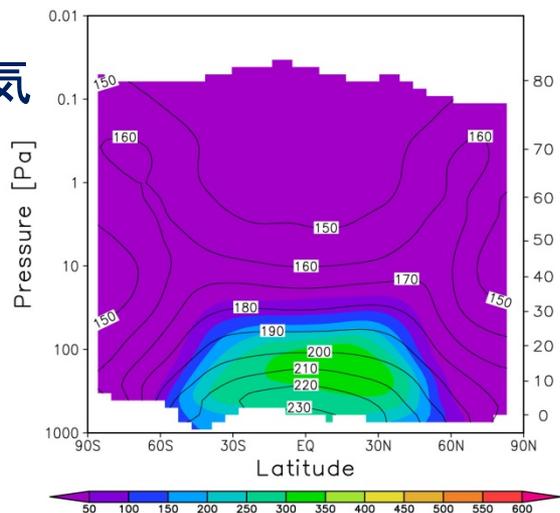
水蒸気と氷雲の鉛直分布[ppm of mass]

等温静止から
20火星年目の
Ls=180°
コンター: 温度

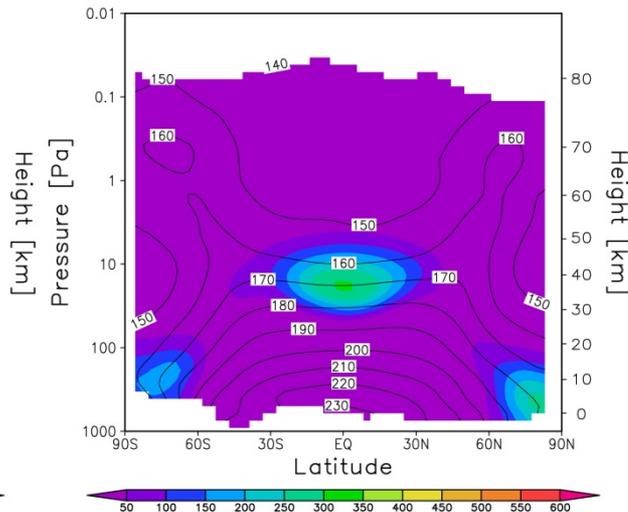
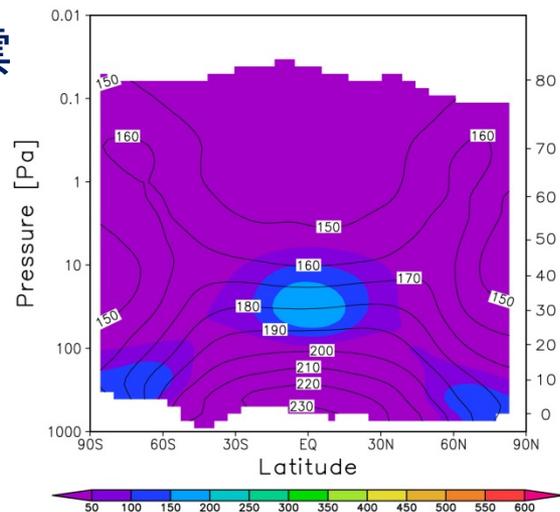
過飽和なし

10倍過飽和

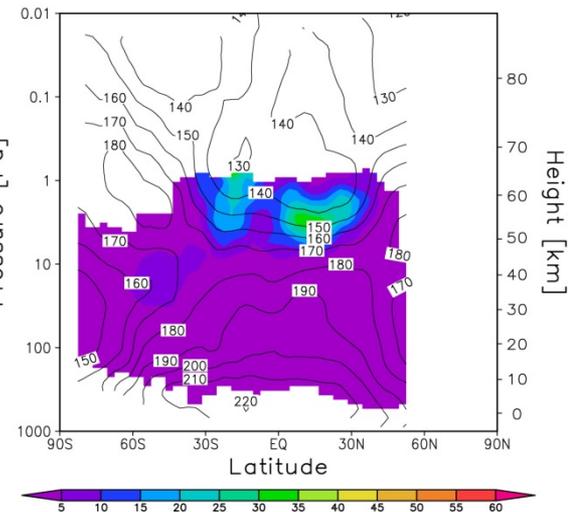
水蒸気



氷雲



参考: MRO-MCS観測
注: シェードのスケールが
モデル結果の10分の1



HDO/H₂O同位体分別(2012年)

SMOW(D/H): 1.56×10^{-4}

Merlivat and Nief [1967]の式

$$\alpha = \frac{(\text{HDO}/\text{H}_2\text{O})_{\text{ice}}}{(\text{HDO}/\text{H}_2\text{O})_{\text{vap}}} = \exp\left(\frac{16289}{T^2} - 9.45 \times 10^{-2}\right)$$

$\alpha=1.72$: 160K

1.51: 180K

1.37: 200K

2種類の理想化された見積もり[Montmessin et al., 2005]

Rapid isotopic Homogenization case (RH):
Appropriate for liquid clouds

$$\alpha = \frac{(M_{hdo}^c + dM_{hdo}) / (M_{h2o}^c + dM_{h2o})}{(M_{hdo}^v - dM_{hdo}) / (M_{h2o}^v - dM_{h2o})}, \quad (1)$$

Rayleigh Distillation case (RD):
Appropriate for ice clouds

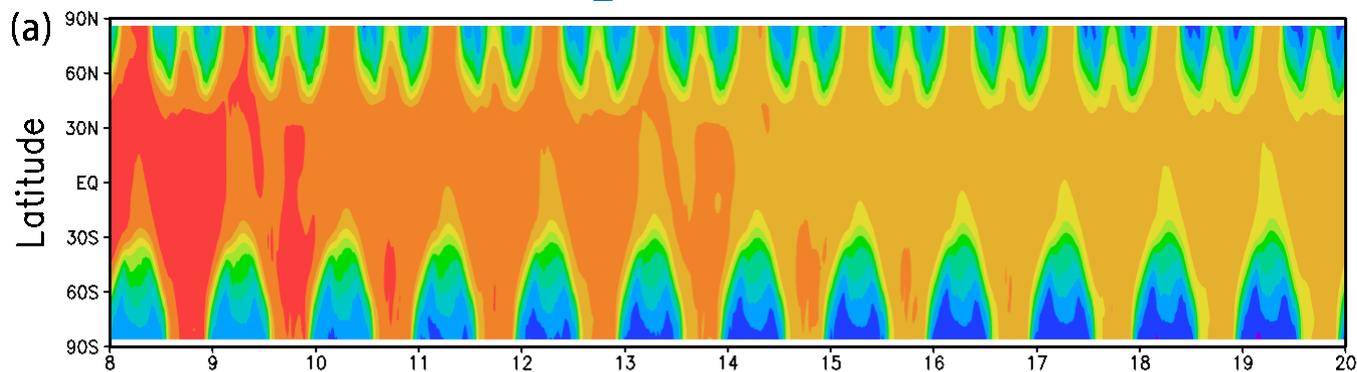
$$\alpha = \frac{dM_{hdo} / dM_{h2o}}{M_{hdo}^v / M_{h2o}^v} \quad (2)$$

大気における分別には(1)を、地表面の氷には(2)を適用

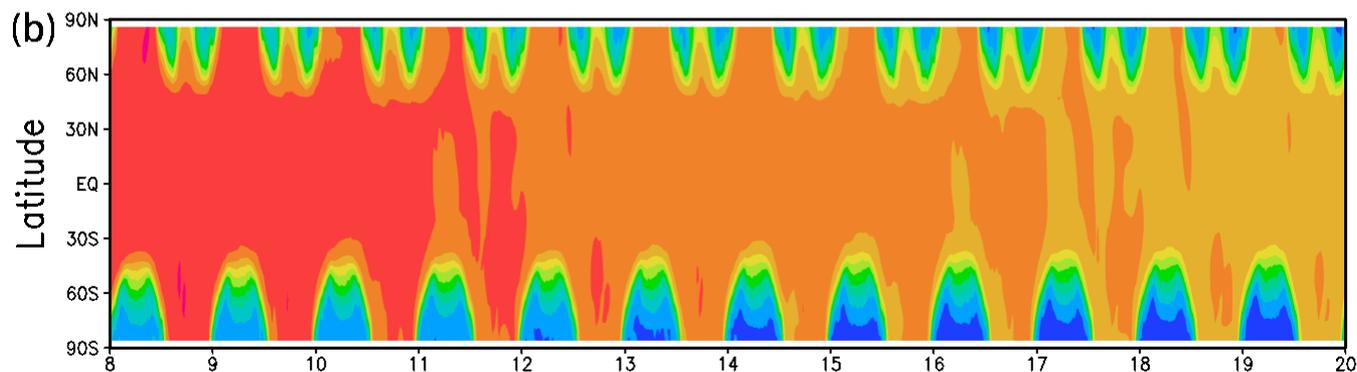
北緯80度以北の氷床におけるHDO/H₂O比を、火星における観測値に基づいてSMOW (地球海洋の平均値)の5.6倍に設定

水蒸気カラム量におけるHDO/H₂O比[vrt. SMOW]

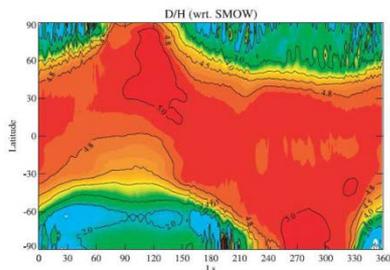
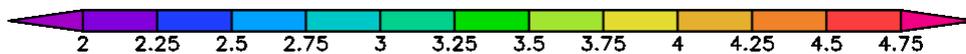
過飽和なし



10倍過飽和



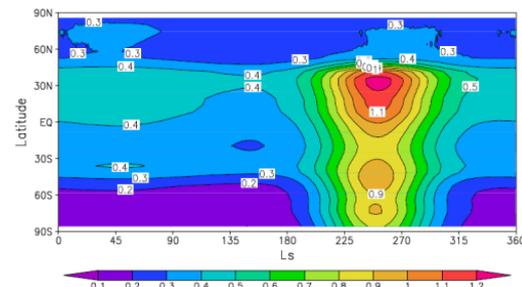
Year



[Montmessin et al., 2005]

Seasonal-latitude change of dust opacity: 0.2~1 in visible

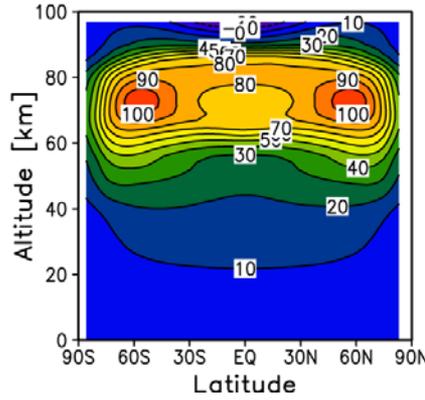
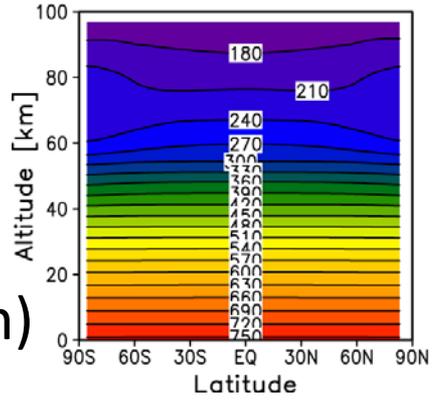
Montmessin et al. [2005]に近い季節・緯度変化は再現されている



金星GCM(東北大着任～)

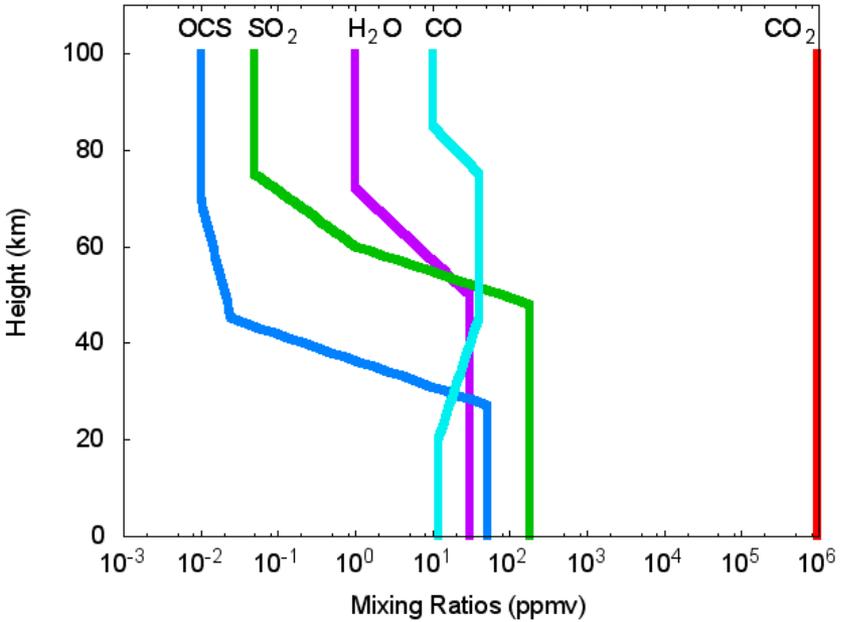
Tohoku/AORI VGCM [池田, 2011]

CCSR/NIES/FRCGC MIROCベース
水平T21(赤道域グリッド間隔~600km)
鉛直52層(モデル上端~95km)

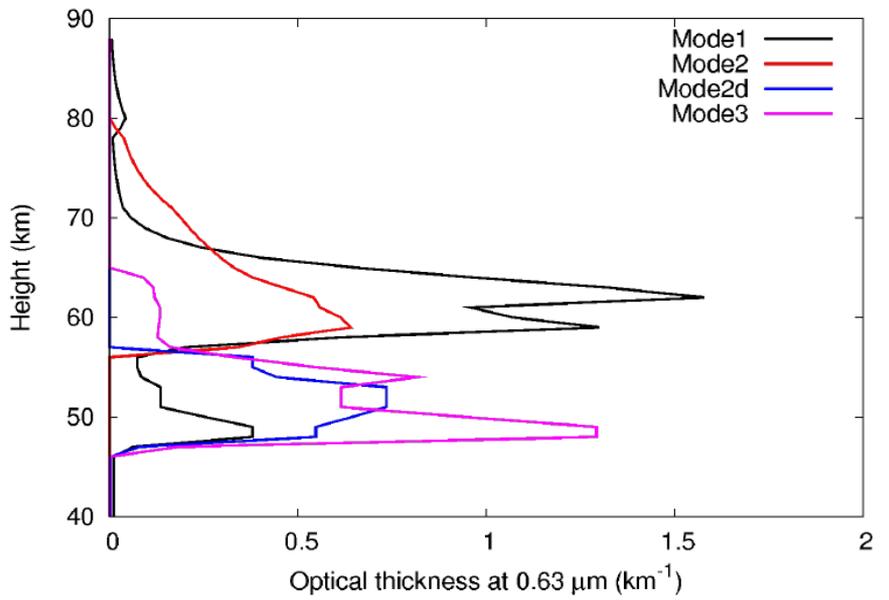


下記の大気分子・雲の鉛直分布に合わせた現実的放射コード+
下層からの重力波強制でスーパーローテーションを再現(右上図)

Component of gases
(from Pollack et al., 1993)



Cloud distributions
(from Crisp, 1986)



Cloud simulations

Step 1: Implementation of condensation/sublimation

- In the next step, the condensation/sublimation process of sulfuric acid cloud are considered using the formula of Kulmala and Laaksonen (1990).
- If the amount of H_2SO_4 is beyond the saturated level, the oversaturated H_2SO_4 immediately becomes the clouds.
- Supersaturation is not considered.

H_2SO_4 vapor ($\times 10^{-11}$ in volume)

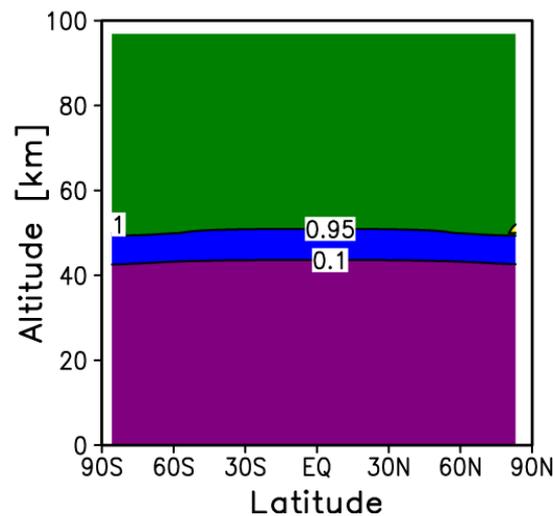
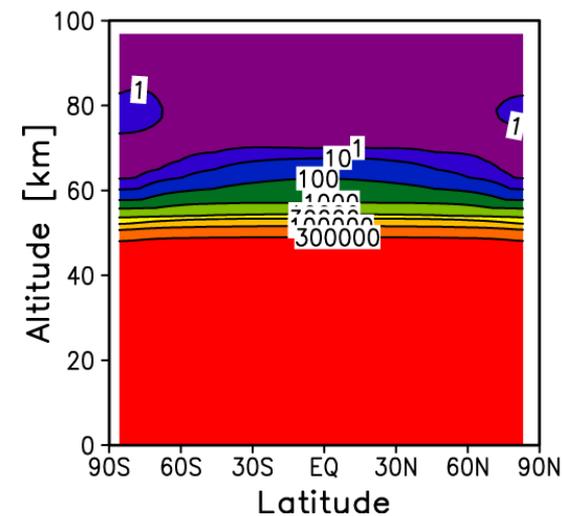
(注:これは非現実的)

Supersaturation rate

Saturation vapor pressure of H_2SO_4
(Kulmala and Laaksonen, 1990)

$${}^0P_a^{\text{sat}} = 101325 \exp\left\{-11.7 + 10156 \left[\frac{1}{360.15} - \frac{1}{T} + \frac{0.38}{545} \left(1 + \ln \frac{360.15}{T} - \frac{360.15}{T} \right) \right]\right\}$$

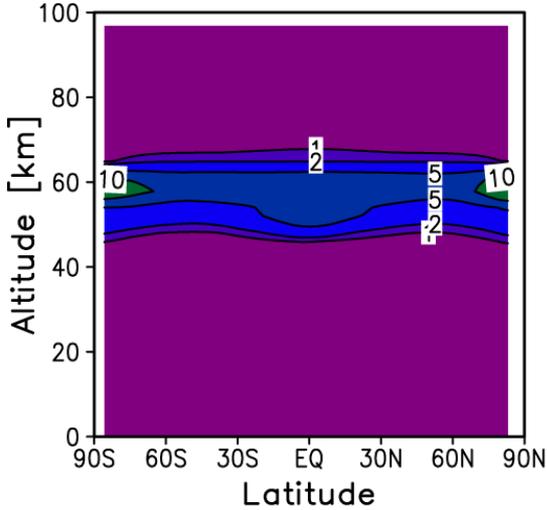
Initial states of H_2SO_4 vapor
and supersaturation rate



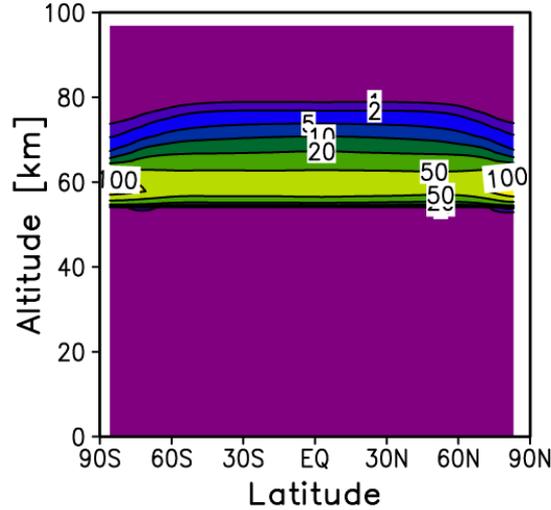
Cloud simulations

Step 1: Implementation of condensation/sublimation

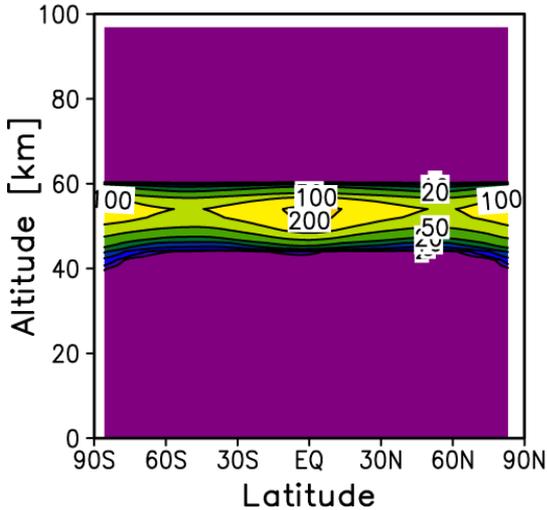
Mode 1



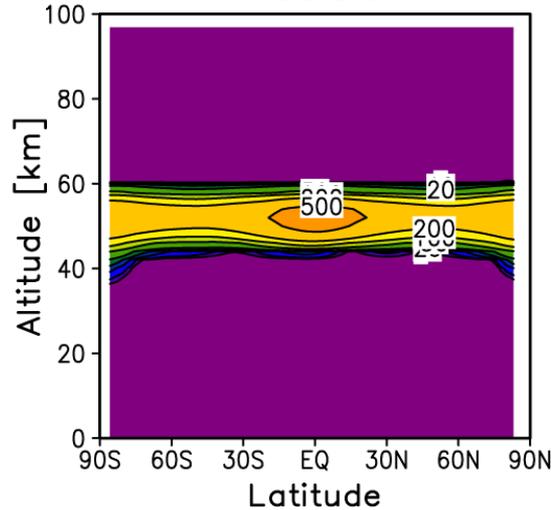
Mode 2



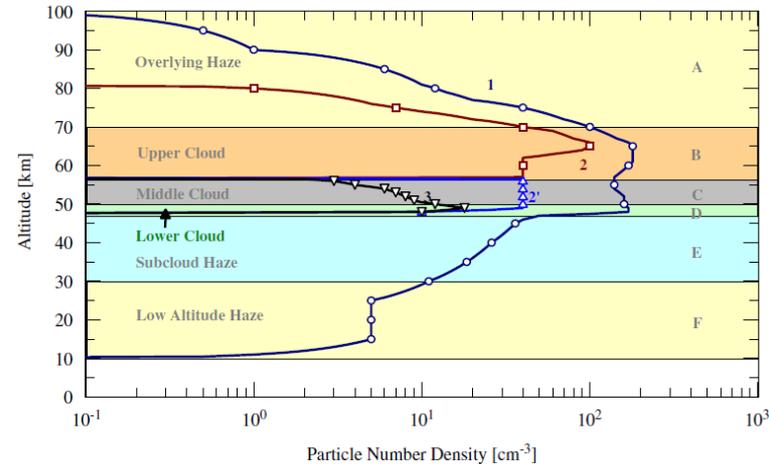
Mode 2'



Mode 3



- Generated clouds are distributed to each mode to keep the ratios at each height.

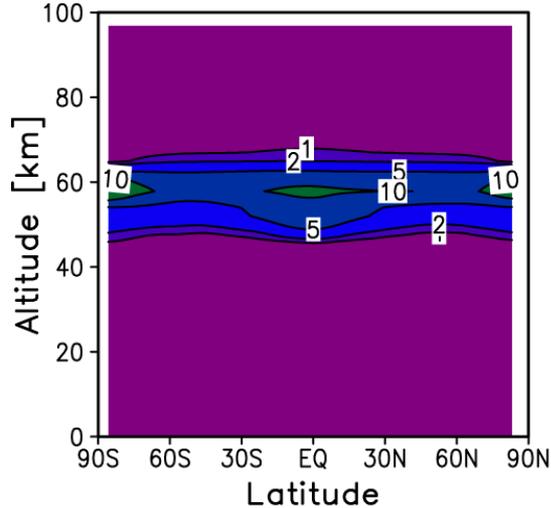


1 Venusian day from the initial state
(volume mixing ratio, $\times 10^{-11}$)

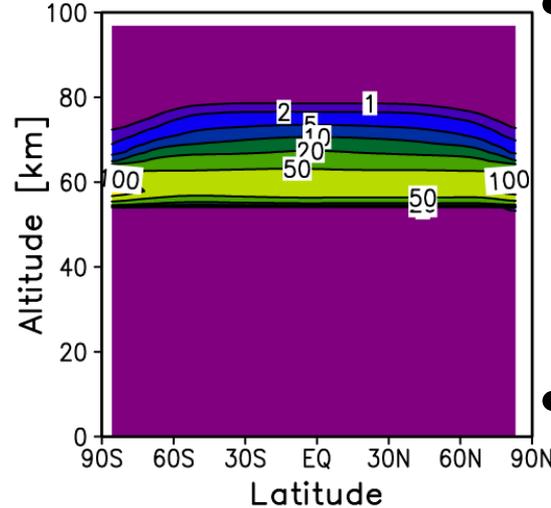
Cloud simulations

Step 1: Implementation of condensation/sublimation

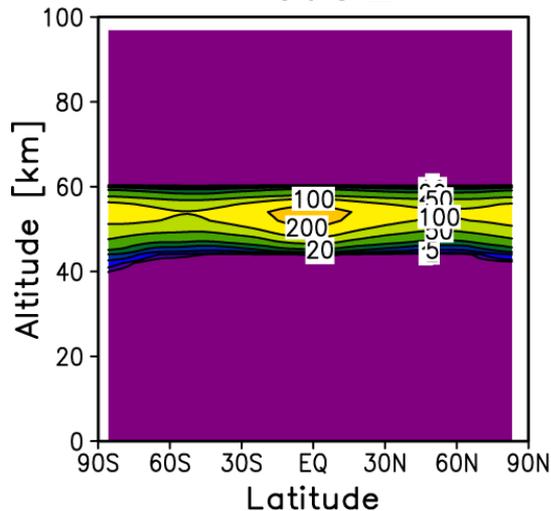
Mode 1



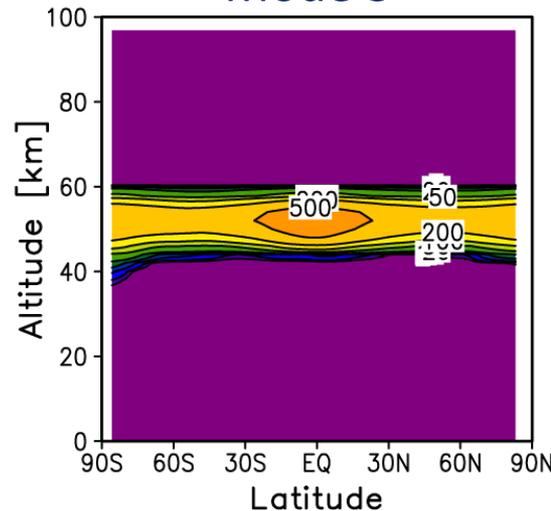
Mode 2



Mode 2'



Mode 3



- Now the simulation achieves the equilibrium of cloud distributions.

(ただし高度は低め: 非現実的な硫酸蒸気分布により)

- But, from the standpoint of cloud simulations, it is not realistic that the clouds of larger mode radius are generated immediately.

3 Venusian days from the initial state
(volume mixing ratio, $\times 10^{-11}$)

Cloud simulations

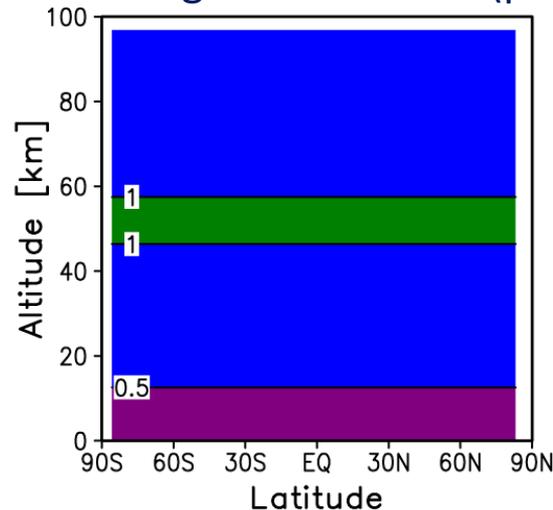
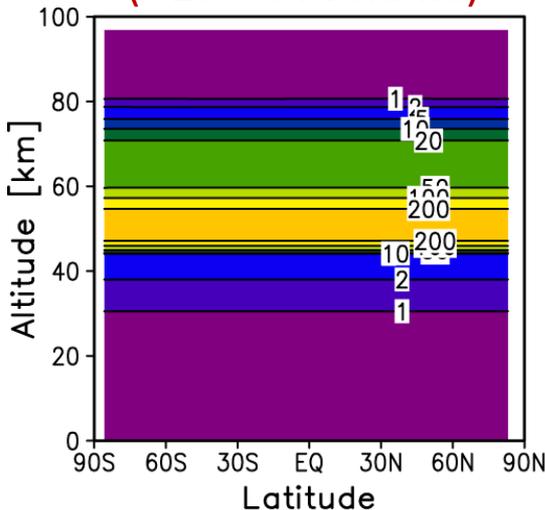
Step 2: Implementation of variable cloud radius

- The number density of clouds at each grid/layer is fixed to the sum of all modes by Haus and Arnold (2010).
- From that and cloud mixing ratio (the sum is treated as one tracer), the average cloud radius at each grid/layer is estimated.
- The cloud sedimentation is treated with the average cloud radius.

Estimated cloud radius
($r_0 = 0.1 \mu\text{m}$) $r = \left(\frac{q_{\text{cld}} \rho_{\text{air}}}{\frac{4}{3} \pi \rho_{\text{cld}} n_{\text{cld}}} + r_0^3 \right)^{\frac{1}{3}}$

Cloud mixing ratio
($\times 10^{-11}$ in volume)

Average cloud radius (μm)



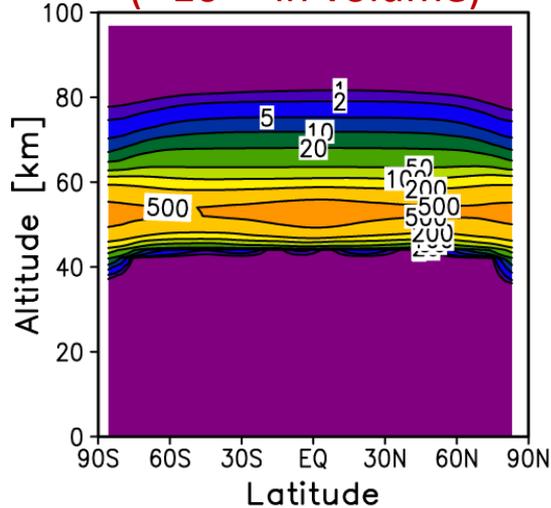
- The condition of cloud formation is same as Step 1 (supersaturation is not considered).

Initial state

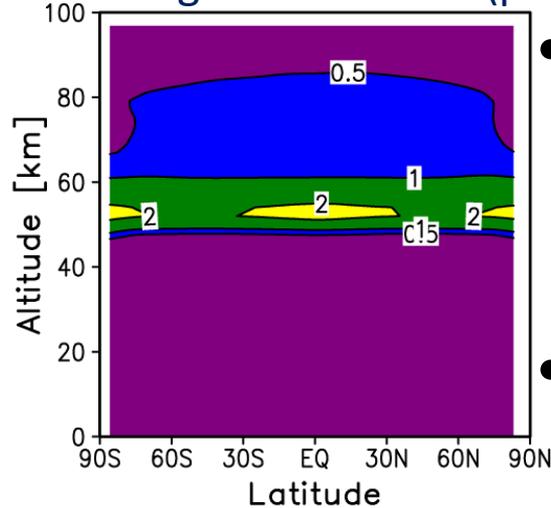
Cloud simulations

Step 2: Implementation of variable cloud radius

Cloud mixing ratio
($\times 10^{-11}$ in volume)



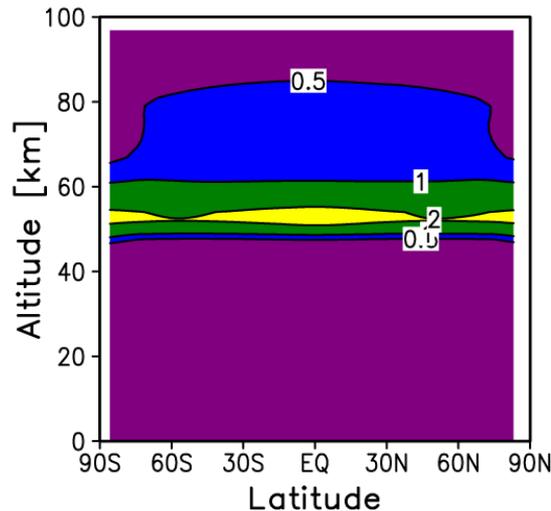
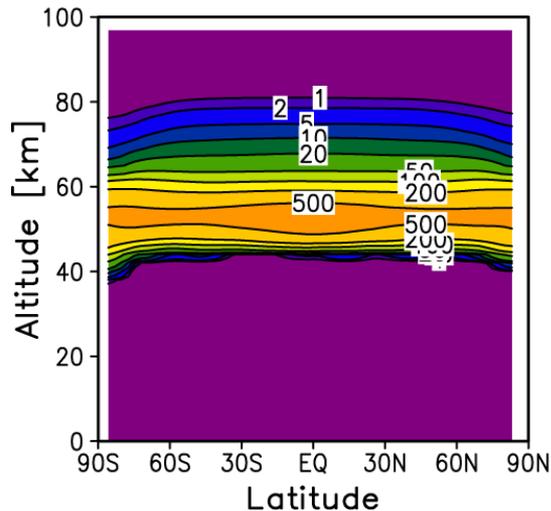
Average cloud radius (μm)



- The results quickly go to the equilibrium state.

(ただし高度は低め: 非現実的な硫酸蒸気分布により)

- But, to reproduce the cloud distributions more realistically, the effect of supersaturation should be potentially important.



After 1 Venusian day (top)
and 3 Venusian days (bottom)
from the initial state

Cloud simulations

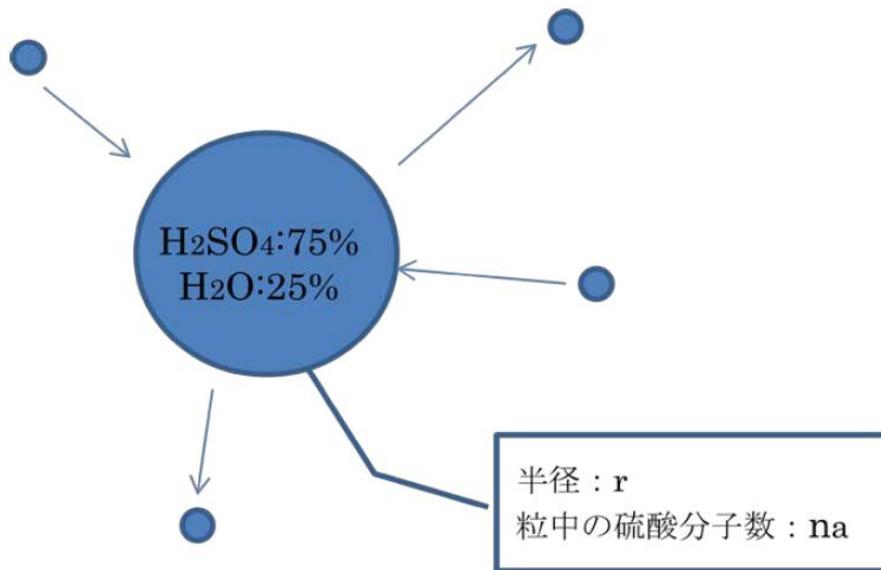
Step 3: Implementation of cloud microphysics

- The change rate of cloud radius is calculated from the cloud microphysics, based on the ideas of Toon and Kiang (1977) and Imamura and Hashimoto (2001).

Change rate of
particle radius

$$\frac{dr}{dt} = \frac{\bar{v}}{4\pi r^2 \chi} \frac{4\pi r D N_a}{1 + \lambda K_n} = \frac{\bar{v} D N_a}{r \chi (1 + \lambda K_n)}$$

$$= \frac{\bar{v} D}{r \chi (1 + \lambda K_n)} \frac{P_a - P_a^{\text{sat}}}{kT}$$



\bar{v} : Average volume of one molecule in the cloud

D: Diffusive coefficient

N_a : Number density of H₂SO₄ in the atmosphere

P_a : Partial pressure of H₂SO₄

P_a^{sat} : H₂SO₄ saturation pressure on the curved surface

χ : Ratio of H₂SO₄ molecule in the cloud

k: Boltzmann coefficient

T: Temperature

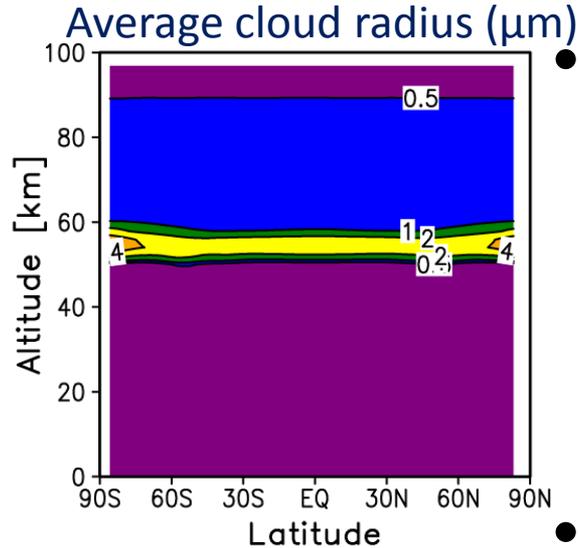
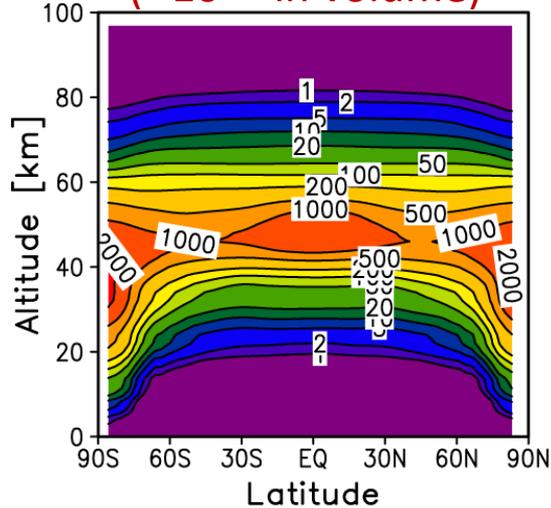
K_n : Knudsen number

λ : Offset of Knudsen number

Cloud simulations

Step 3: Implementation of cloud microphysics

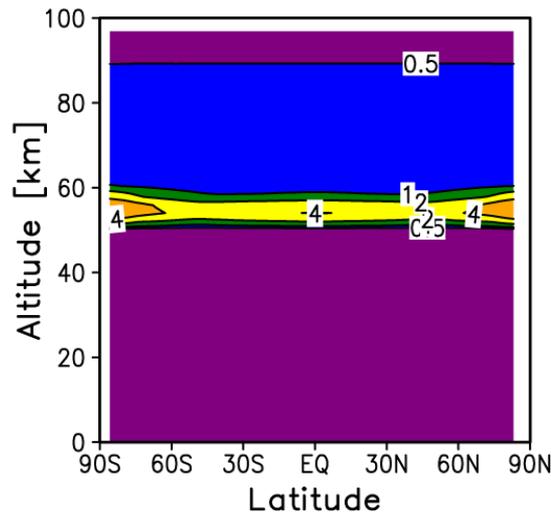
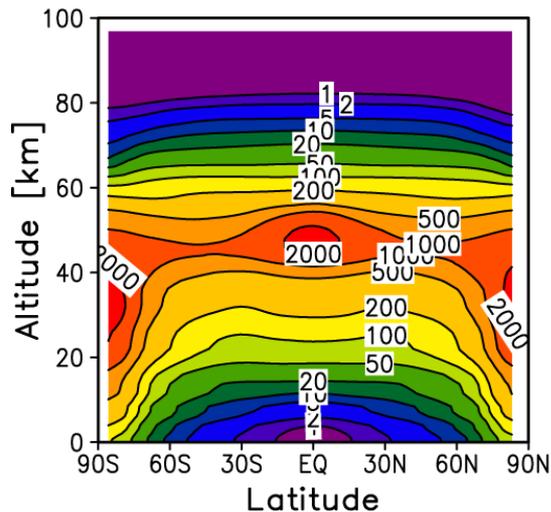
Cloud mixing ratio
($\times 10^{-11}$ in volume)



- As the clear difference from Step 2, the clouds in polar regions tend to be thicker to lower altitude with larger radius than at the equator.

- But the clouds reach to surface without evaporating, which looks unrealistic.

After 1 Venusian day (top)
and 3 Venusian days (bottom)
from the initial state

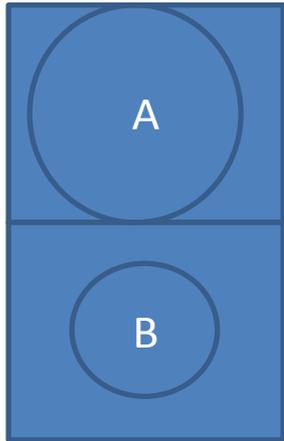


Cloud simulations

Step 3: Implementation of cloud microphysics

Why the results...?

- In the current setting the particle size is defined to each grid, not to the clouds themselves.



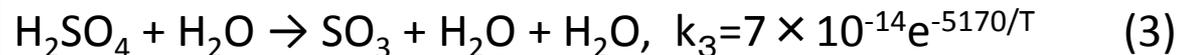
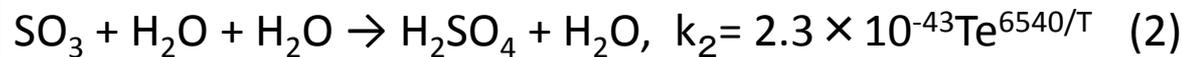
When the particle size of Grid A becomes larger, only the cloud mass goes down to Grid B by gravitational sedimentation, without changing the cloud sizes in both grids.

→ Troubles of unreality can be occurred.

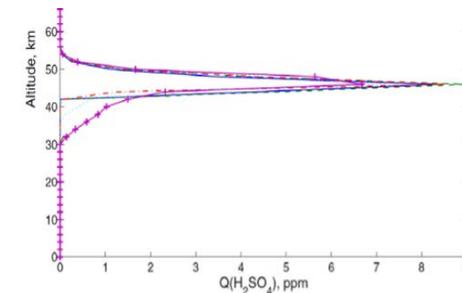
- To improve that, the information of particle sizes should be moved together with the advection/sedimentation of cloud mass. → **Step 4 (under consideration)**

Cloud simulations (+ Chemistry)

反応式 [Krasnopolsky, 2013, 2014]



硫酸蒸気鉛直分布も
現実的なものに修正



(触媒MはCO₂とした)

高度47-77kmに(1)、(2)を、地表面から47kmに(2)、(3)、(4)を導入

(1)-(4)式を(5)のように表し、数密度を気体の状態方程式で質量混合比に変換し、各グリッドで計算を行った

$$\frac{d}{dt} [\text{SO}_3] = k_1 [\text{SO}_2][\text{O}][\text{M}] \quad (5)$$

k_i : 反応係数 ($\text{cm}^3 \text{s}^{-1}$, $\text{cm}^6 \text{s}^{-1}$)

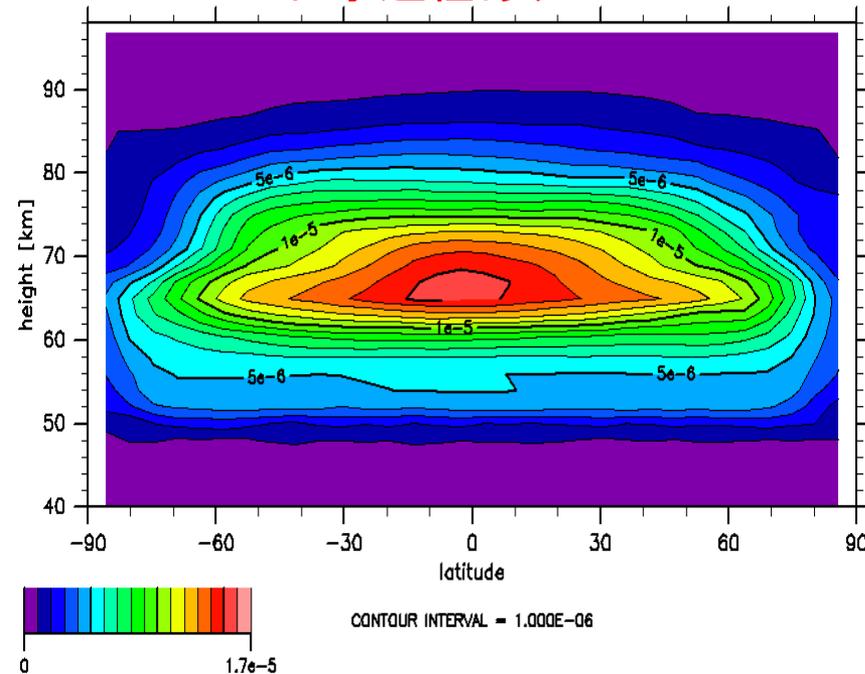
$[\text{M}]$: 数密度 (cm^{-3})

Cloud simulations (+ Chemistry)

Itoh et al. (2014)

結果(雲量)

化学過程あり



化学過程なし

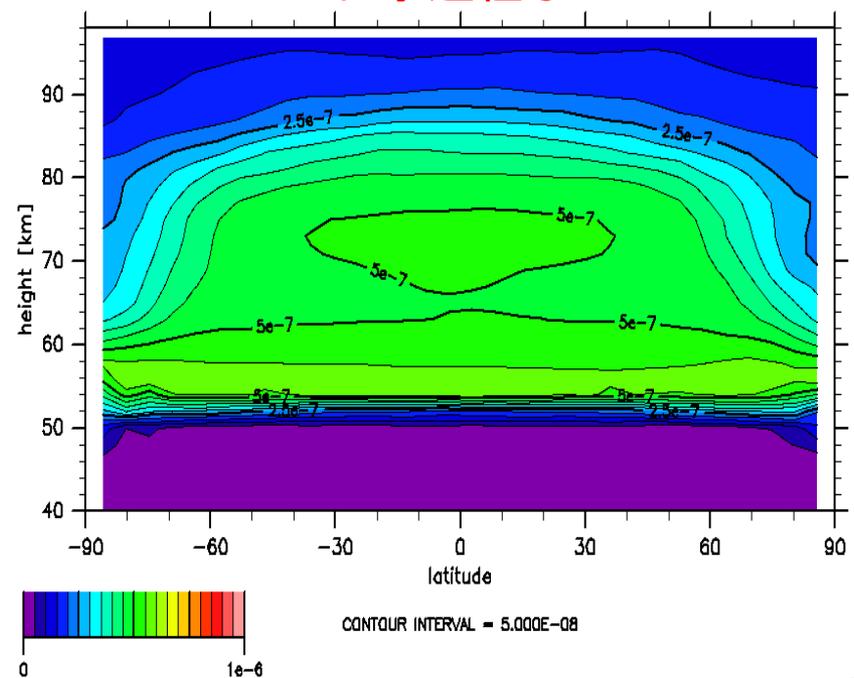


図5: 雲量(質量混合比)の緯度-高度分布図

- ✓ 化学過程を導入したものはしていないものに比べて2ケタほど大きくなった

Cloud simulations (+ Chemistry)

結果(雲量)

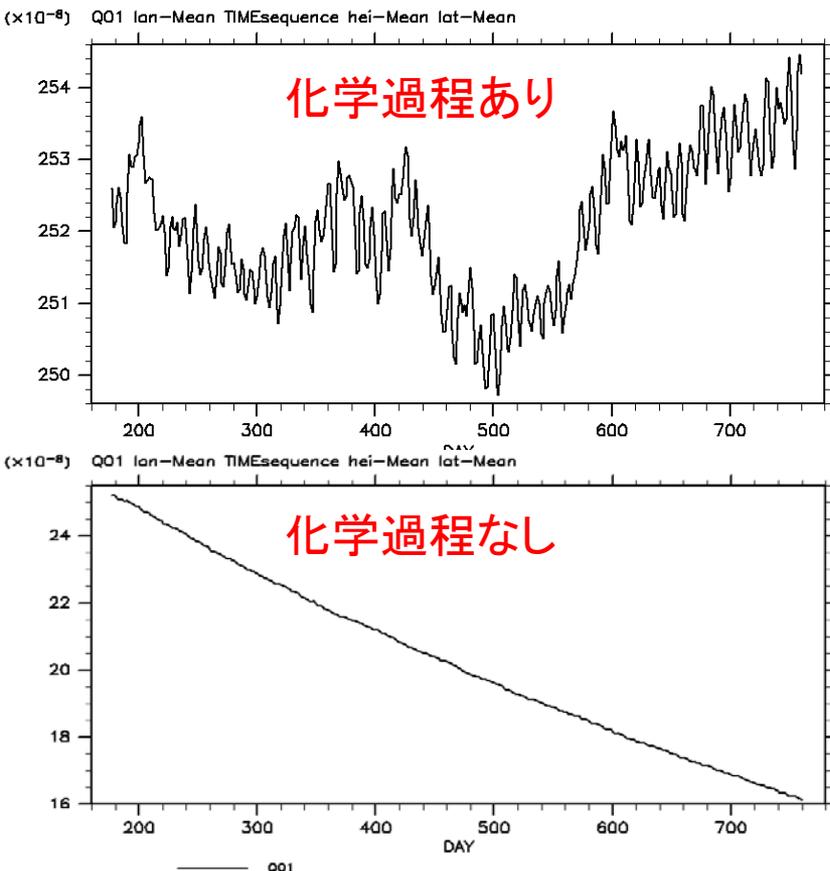


図6: 雲量の全球平均の時間変化
(横軸は地球日)

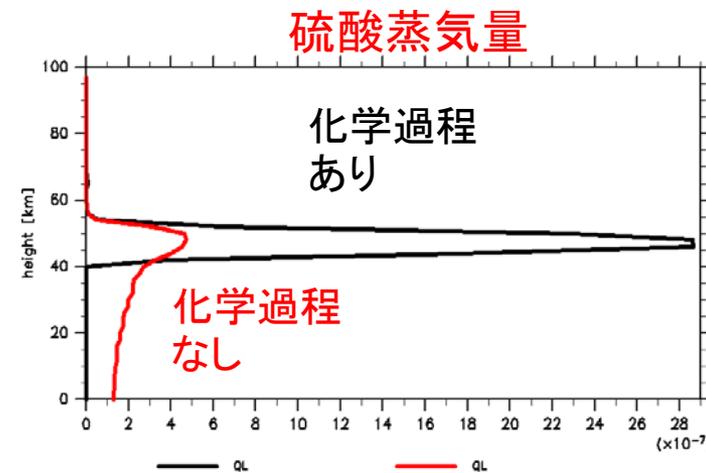


図7: 硫酸蒸気量を水平方向に平均した鉛直プロファイル

- ✓ 化学過程を導入したものがほぼ平衡に達している
- ✓ 導入していないものは平衡に達せず減少し続けている
- ✓ 硫酸蒸気量が下層にたまってしまっている

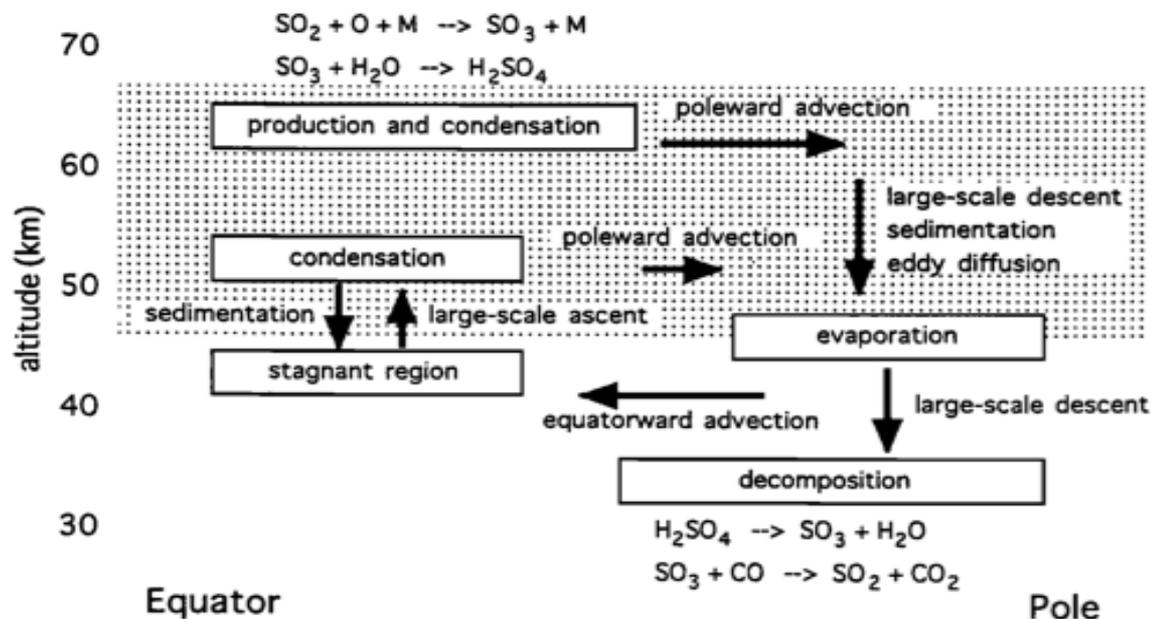


化学過程を考慮しないと硫酸雲の生成・消失サイクルを維持できない

Cloud simulations (+ Chemistry)

結果(まとめ)

化学過程を導入したモデルにおいて、
 雲層の上部(高度約67-80km)でSO₂の消失および硫酸蒸気量の生成、
 雲層の下部(高度約50km付近)でのSO₂の生成および硫酸蒸気量の消失という結果 → Imamura and Hashimoto [1998] の2次元モデルにより示唆された硫酸蒸気サイクル(下図)と整合的



今後の野望

火星GCM

水循環の改良(金星で手をつけた雲粒成長過程の導入など)
氷雲放射効果の導入

表層過程(地下～大気間の水の出入りなど)

大気光化学過程の導入(Lefevre et al. [2004]をベースに検討中)

→DRAMATIC = Dynamics, RAdiation, MAterial Transport,
Isotopomer and Chemistry になる予定

大気散逸過程とのカップリング、気候変動へのアプローチ

データ同化(withスウェーデン・チャルマス工科大)

国内外の探査ミッションへの貢献

金星GCM

放射・物質循環など全体的により洗練されたモデルに
観測との比較、「あかつき」への貢献