Overestimated acceleration of the Brewer-Dobson circulation due to stratospheric cooling

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Stratospheric Dynamics

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Temperature changes affect atmospheric structure

Hypsometric equation (assuming dry air and hydrostatic balance):

$$\phi(p_2) - \phi(p_1) = g_0 \cdot (Z(p_2) - Z(p_1)) = R \int_{p_2}^{p_1} T d \ln p \quad (1)$$

$\phi(p)$: geopotential of pressure levels
$Z(p)$: geopotential height of pressure levels

$\Rightarrow$ temperature trends lead to changes in height of pressure levels

as the troposphere warms, it expands
as the stratosphere cools, it shrinks

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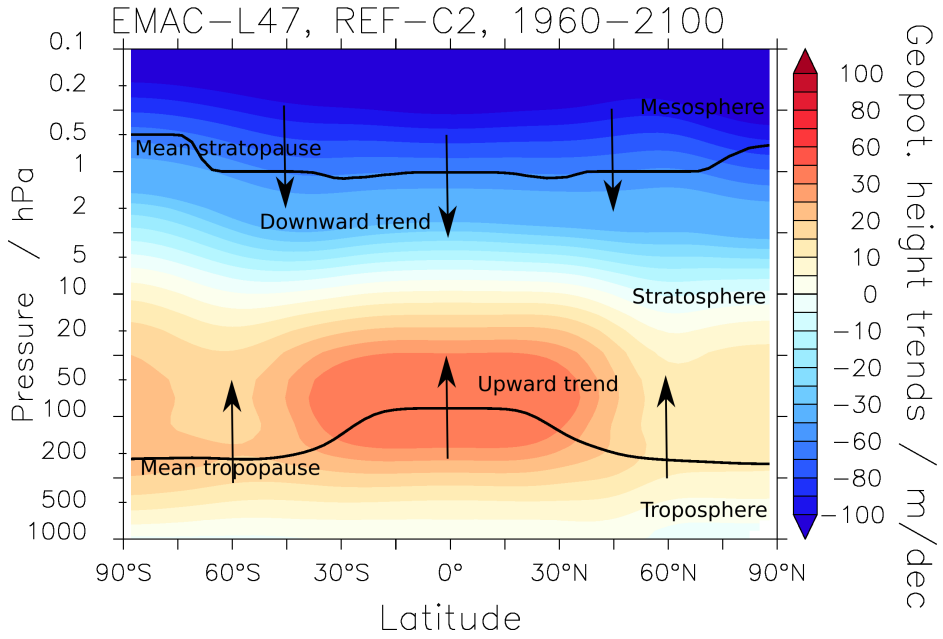
Stratospheric shrinkage

Stratospheric cooling leads to

- reduction of upward shift with altitude in stratosphere
- downward shift in upper stratosphere and mesosphere
  (Lübken et al., 2009; Berger and Lübken, 2011; Laštovička et al., 2006)



The stratospheric thickness decreases ("stratospheric shrinkage")
(Šácha et al., 2019; Pišoft et al., 2020)

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Stratospheric shrinkage

- The upper atmosphere is generally cooling and contracting affecting the height of ionospheric layers and also spacebased technological systems *(Laštovička et al., 2006)*.
- Stratospheric shrinking can cause temperature changes in the mesosphere *(Lübken et al., 2009)*.
- Trends at geometric heights are larger than trends at pressure levels due to atmos. shrinking *(Berger and Lübken, 2011; Šácha et al., 2019)*.

CCMI-1 REF-C2 model simulations

Model	Distance betw. 1 and 100 hPa / m	Trend of distance / m/dec.	Trend of distance / %/dec.
WACCM	30915	-62	-0.20
ULAQ	31112	-76	-0.25
SOCOLv3	30873	-68	-0.22
NIWA-UKCA	31026	-68	-0.22
MRI	31414	-70	-0.23
GEOSCCM	31096	-67	-0.22
EMAC-L90	30746	-72	-0.24
EMAC-L47	30827	-73	-0.24
CMAM	31359	-75	-0.24
CCSRNIES	30795	-66	-0.22
ACCESS	31081	-69	-0.23
MMM	**31022**	**-70**	**-0.23**
σ(MMM)	210	4	0.01

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The residual vertical velocity ω^*, w*

- The residual vertical velocity (commonly at 70 hPa) is often used as a proxy for the strength of the residual circulation (the BDC).

  (see e.g. Butchart et al., 2010; Abalos et al., 2015; Dietmüller et al., 2018)

- Due to their design, most models deliver ω^* in $Pa \cdot s^{-1}$, which traditionally is then converted to w* in $m \cdot s^{-1}$ for consistency among models.

- For that, the CCMI-1 data request *(Hegglin and Lamarque, 2015)*: provides the log-pressure equation

$$\omega^* \approx w^* \cdot \frac{-p}{H}, \tag{2}$$

which includes the constant scale height $H = 6950\,\mathrm{m}$

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Residual vertical velocity w*

Some model groups (EMAC, SOCOL) used the more general geometric equation for transformation:

$$\omega^* \approx \frac{dp}{dt} = \frac{\partial z}{\partial t}\frac{\partial p}{\partial z} = w^* \cdot \frac{-p \cdot g}{R \cdot T} \tag{3}$$

⇒ Climatological Δw^* at 70 hPa in EMAC: 17% *(Dietmüller et al., 2018)*

But what does that mean for the trends?

$$H(T) = \frac{R \cdot T(t)}{g} \tag{4}$$

⇒ Stratospheric scale height $H(T)$ significantly decreases over time due to stratospheric cooling and this is not captured in log-pressure system.

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⇒ Assessing the trend differences of

$$w_H^*(t) = \omega^*(t) \cdot \frac{H}{-p} \tag{5}$$

$$w_T^*(t) = \omega^*(t) \cdot \frac{R \cdot T(t)}{-p \cdot g} \tag{6}$$

yields the impact of stratospheric cooling "on the BDC"

For EMAC-L90, EMAC-L47, SOCOL, NIWA[1] and ACCESS[1] we calculate

$$w_H^* = w_T^* \cdot \frac{H \cdot g}{R \cdot T} \tag{7}$$

for the other models (MRI, GEOSCCM, CMAM, NIES and WACCM)

$$w_T^* = w_H^* \cdot \frac{R \cdot T}{H \cdot g}. \tag{8}$$

[1] because w* is calculated directly in geometric $m \cdot s^{-1}$ in these models

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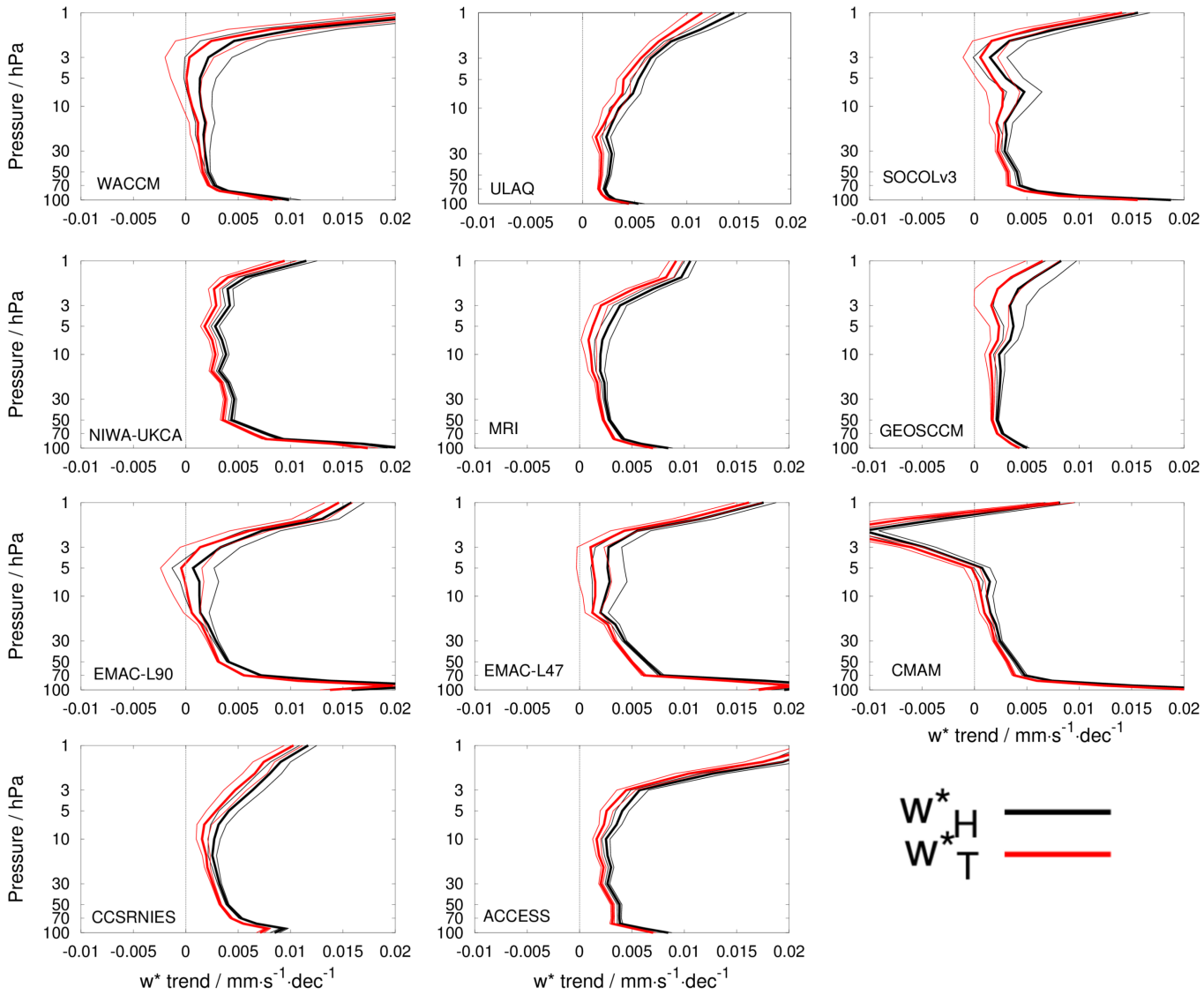
w* linear trend profiles 1960-2100 (CCMI REF-C2)



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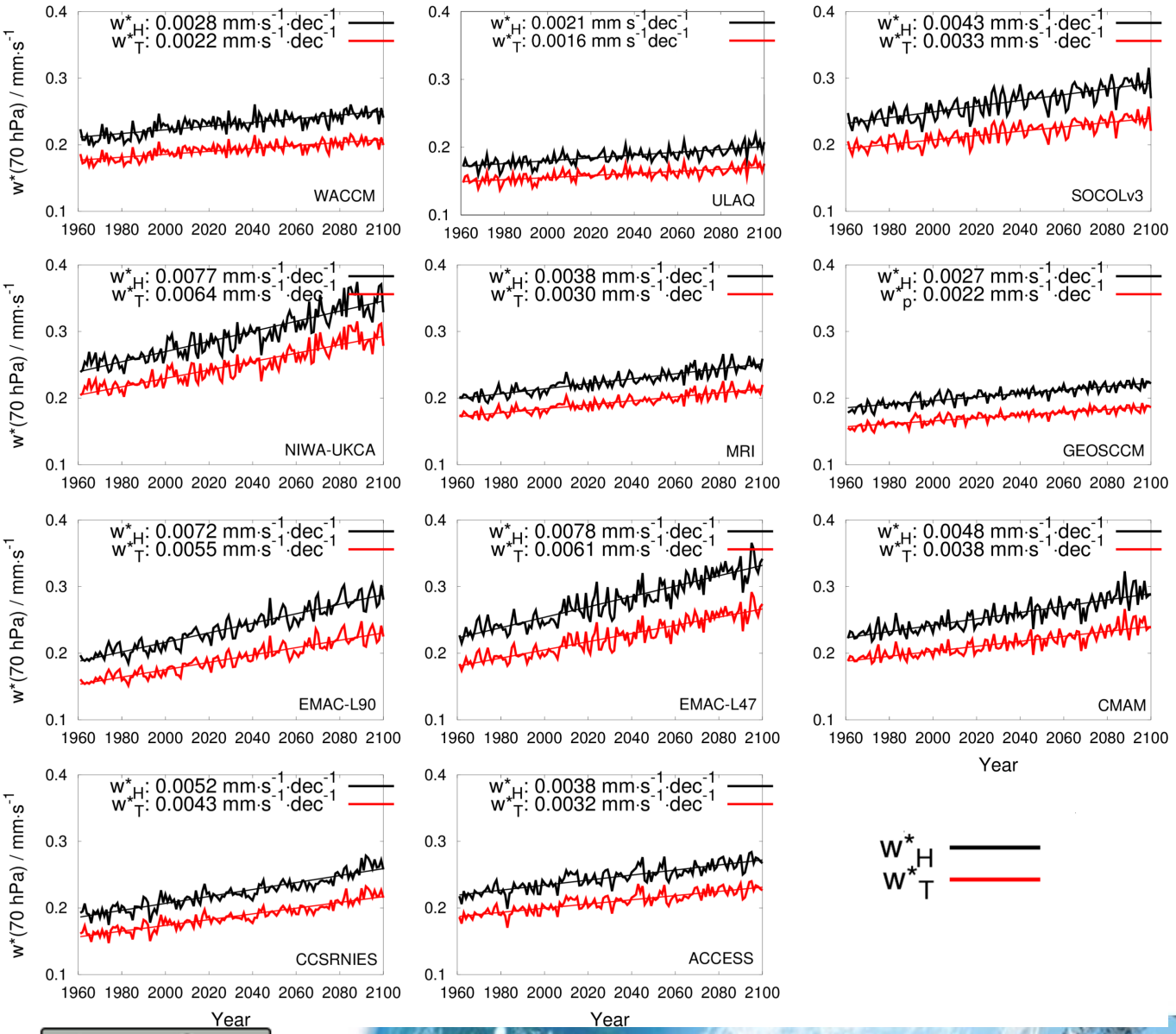
w* time series at 70 hPa 1960-2100 (CCMI REF-C2)



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w^* linear trend at 70 hPa (CCMI REF-C2 1960-2100)

Model	w^*_H mm/s/dec	w^*_T mm/s/dec	Δw^* mm/s/dec	Δw^* %
WACCM	0.0028	0.0022	0.0006	20.7
ULAQ	0.0021	0.0016	0.0005	26.0
SOCOLv3	0.0043	0.0033	0.0011	24.8
NIWA	0.0077	0.0063	0.0013	17.0
MRI	0.0038	0.0030	0.0008	20.1
GEOSCCM	0.0027	0.0022	0.0005	19.3
EMAC-L90	0.0072	0.0055	0.0016	22.4
EMAC-L47	0.0078	0.0061	0.0017	21.2
CMAM	0.0048	0.0038	0.0010	21.4
CCSRNIES	0.0053	0.0043	0.0009	17.4
ACCESS	0.0038	0.0032	0.0006	16.7
MMM	**0.0047**	**0.0038**	**0.0010**	**20.7**
σ(MMM)	0.0019	0.0016	0.0004	2.9

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w^* trend differences

- w^*_H trend systematically larger than w^*_T trend due to stratospheric cooling and Eqs. 7 and 8.

  $\leftarrow$ because units of vertical velocities are different although written identically - w^*_T in geopotential $m \cdot s^{-1}$, w^*_H in log-pressure $m \cdot s^{-1}$.

- Distance between pressure levels decreases in geometric meters, but remains constant in log-pressure meters.

  $\leftarrow$ Difference between w^* trends can be misinterpreted as additional acceleration of the advective BDC part.

- At 70 hPa, $\sim$ 20% of the BDC acceleration in climate projection simulations is due to neglecting stratospheric cooling, if analysed with log-pressure formulas.

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Residual streamfunction and mass flux

The residual streamfunction includes w* in its definition:

$$\psi^*(\phi) = \int 2\pi a^2 \cdot cos(\phi) \cdot \rho w^* d\phi \qquad (9)$$

a: mean Earth's radius, ϕ: latitude, ρ: density

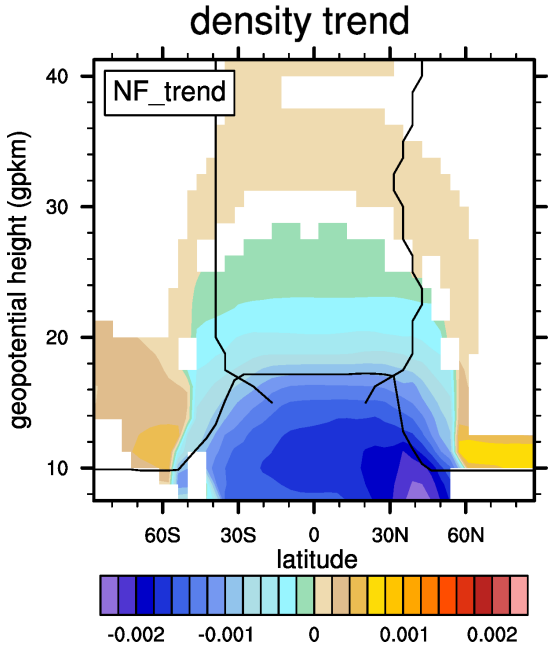
ρ either with constant scale height in log-pressure system:

$$\rho_c = \rho_0 \cdot exp(-\frac{z}{H}) = \frac{p}{g \cdot H} \qquad (10)$$

($\rho_0 = \frac{p_0}{R \cdot T_0}$; z: log-pressure height)

or using the ideal gas law:

$$\rho_T(T) = \frac{p}{R \cdot T}. \qquad (11)$$



(Šácha et al., 2019)

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Residual streamfunction and mass flux

Scaling of w* with ρ for streamfunction calculation:

$$\mathrm{w}^*_T(T) \cdot \rho_T(T) = \omega^* \frac{RT}{-pg} \cdot \frac{p}{RT} = \omega^* \cdot \left(\frac{1}{-g}\right) \tag{12}$$

$$\mathrm{w}^*_H \cdot \rho_c = \omega^* \frac{H}{-p} \cdot \frac{p}{gH} = \omega^* \cdot \left(\frac{1}{-g}\right) \tag{13}$$

$\Rightarrow$ The density effect cancels out the w* effect ($\Rightarrow \psi^*_T = \psi^*_H$)

$\Rightarrow$ only quantities that include metres (in vertical direction) in their units are affected (...if the variables are used consistently)

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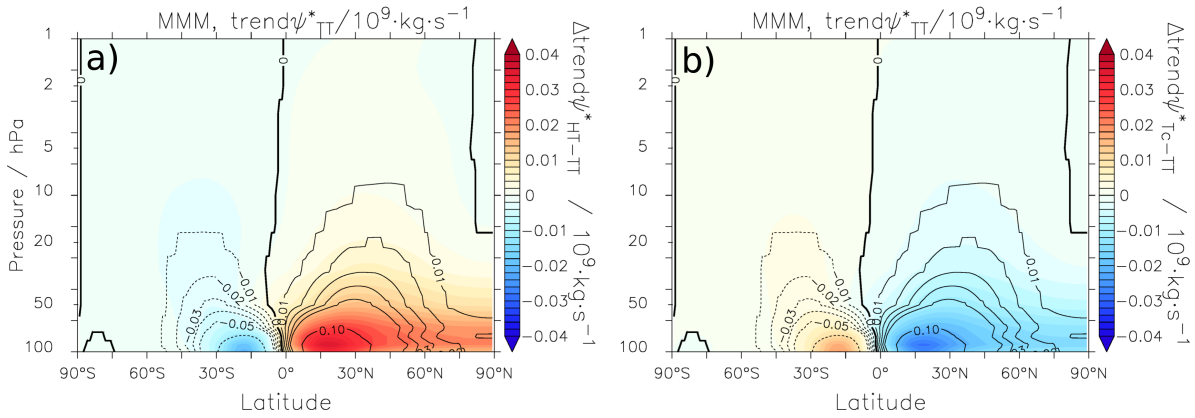
Residual streamfunction and mass flux

Inconsistent usage of w* and ρ (i.e. $w^*_H \cdot \rho_T$ to yield ψ^*_{HT}, or $w^*_T \cdot \rho_c$ to yield ψ^*_{Tc}) leads to streamfunction trend (and climatology) differences:



For the tropical upward mass flux

$$F(p,\phi) = max(\psi^*(p,\phi)) - min(\psi^*(p,\phi)) \tag{14}$$

that leads to differences of 28.5% for F_{HT} and -21.5% for F_{Tc} at 70 hPa in MMM (see additional slides for individual models).

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Summary and conclusions

Stratospheric shrinkage...

- happens due to stratospheric cooling
- leads to $\sim$70 $mdec^{-1}$ less stratosphere, robustly simulated by CCMs
- leads to trend differences if analysed on different coordinates

The scale height...

- decreases due to the reduction of pressure level distance
- trend distorts the trends of w^* if it is assumed to be constant

The w^* trend...

- is enhanced through the constant scale height assumption, i.e. through stratospheric shrinkage
- bias at 70 hPa is $\sim$20% consistently among the CCMI-1 REF-C2 simulations if analysed with log-pressure formulas

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Summary and conclusions

⇒ Previous studies may have overestimated the BDC acceleration (by means of 70 hPa w^* analyses) by 20%.

- Other diagnostics can be affected too, but streamfunction is not because H and ρ trends cancel each other out (if calculated consistently)
- Most multi-model initiatives (CCMVal, CCMI, CMIP) use the log-pressure formulas for these diagnostics,
  ⇒ these data do not consider the effect of varying geopotential height of pressure levels and should therefore not be used for trend studies

We call to...

- be cautious when making trend studies of dynamic variables (and tracer trend comparisons in different coordinate systems)
- change the data requests (CCMI-2) of the multi-model assessments

THANK YOU

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Additional slides

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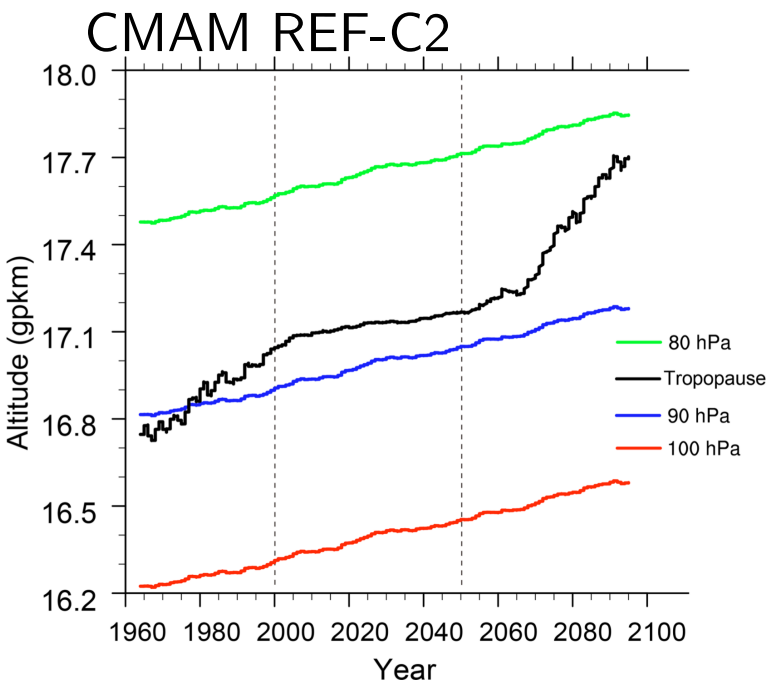
Pressure level trend



(Šácha et al., 2019)

- Due to the reversal in signs of ODSs and O_3 trends, the year 2000 marks a change in stratospheric dynamics
- This impacts the tropopause height trend, but not really the pressure level trend

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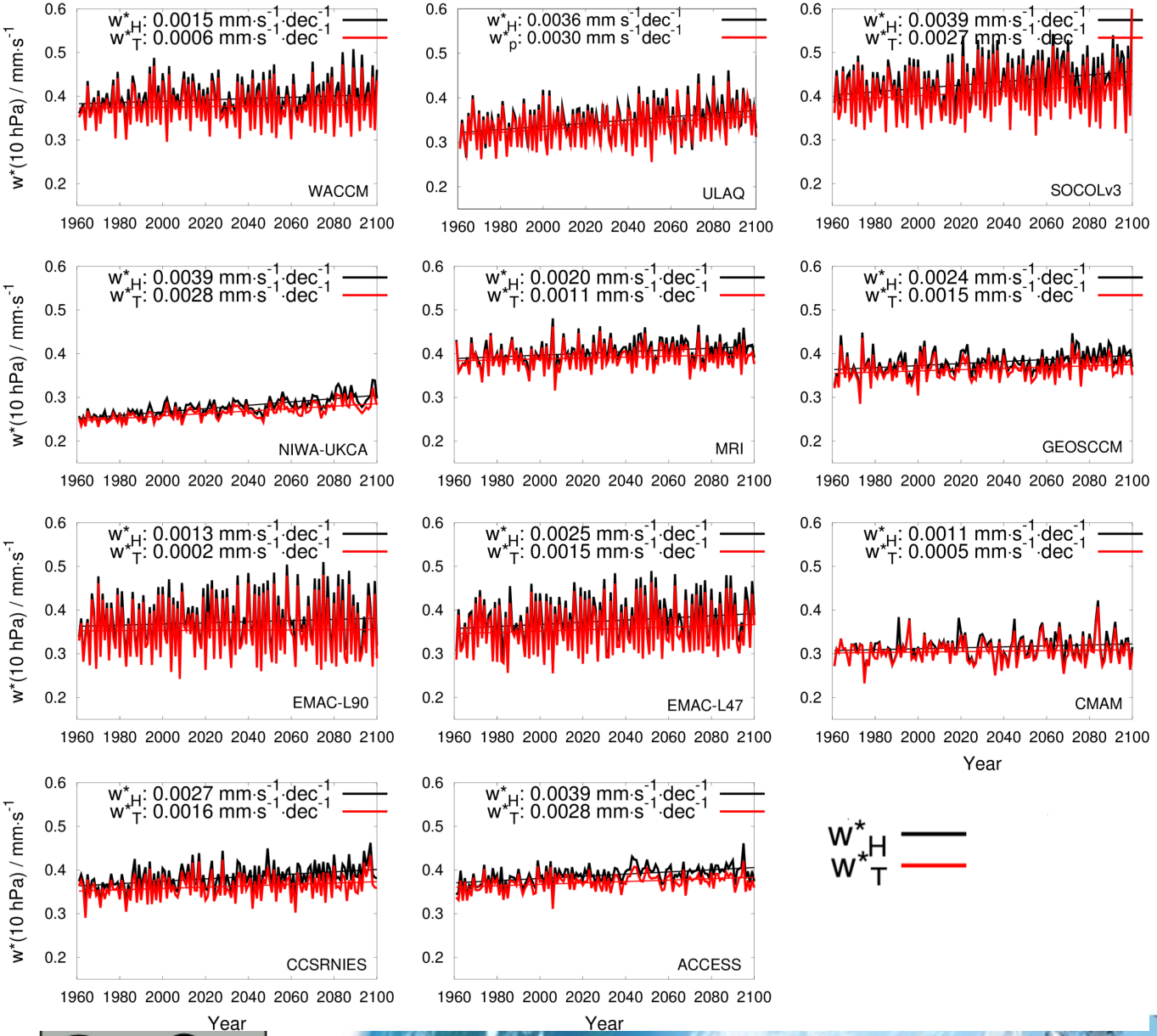
w* time series at 10 hPa (CCMI REF-C2 1960-2100)



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w^* trend 10 hPa (CCMI REF-C2 1960-2100)

Model	w^*_H mm/s/dec	w^*_T mm/s/dec	Δw^* mm/s/dec	Δw^* %
WACCM	0.0015	0.0006	0.0009	60.0
ULAQ	0.0036	0.0030	0.0006	17.1
SOCOLv3	0.0039	0.0027	0.0012	31.6
NIWA	0.0039	0.0028	0.0010	26.5
MRI	0.0020	0.0011	0.0009	45.7
GEOSCCM	0.0024	0.0015	0.0009	37.3
EMAC-L90	0.0013	0.0002	0.0011	84.7
EMAC-L47	0.0025	0.0015	0.0010	39.2
CMAM	0.0011	0.0005	0.0006	51.2
CCSRNIES	0.0027	0.0016	0.0012	43.0
ACCESS	0.0025	0.0016	0.0009	35.3
MMM	**0.0025**	**0.0016**	**0.0009**	**42.9**
σ(MMM)	0.0009	0.0009	0.0002	17.3

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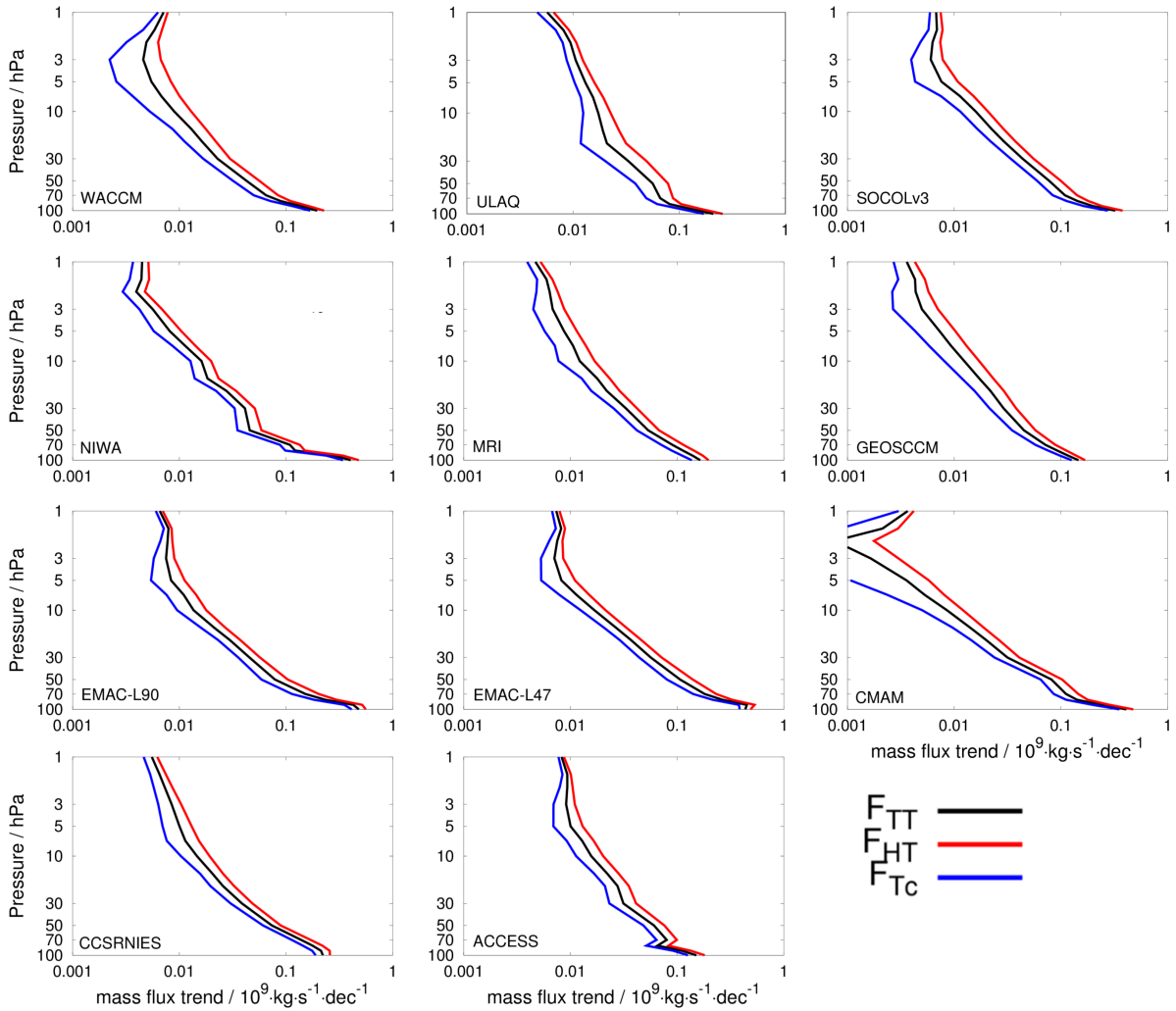
Tropical mass flux trend profiles (1960-2100, CCMI-1 REF-C2)



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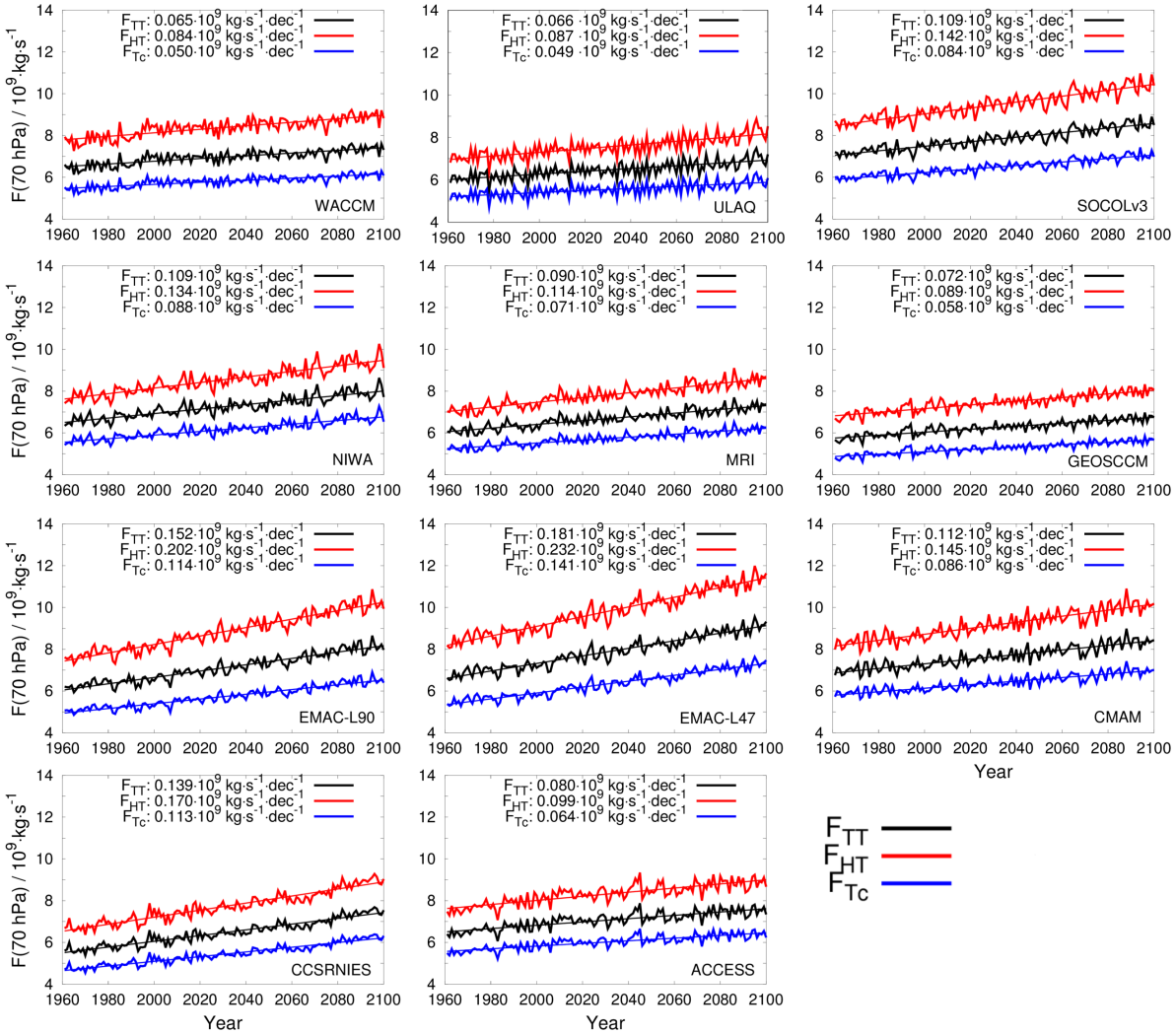
Time series of annual mean tropical mass flux estimates



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Trends of mass flux estimates and differences

Model Model	F_{TT} tr. / $10^9 \cdot kg \cdot s^{-1} dec^{-1}$	F_{HT} tr. / $10^9 \cdot kg \cdot s^{-1} dec^{-1}$	F_{Tc} tr. / $10^9 \cdot kg \cdot s^{-1} dec^{-1}$	ΔF_{HT-TT} tr. / $10^9 \cdot kg \cdot s^{-1} dec^{-1}$	ΔF_{Tc-TT} tr./ $10^9 \cdot kg \cdot s^{-1} dec^{-1}$	ΔF_{HT-TT} trend / %	ΔF_{Tc-TT} trend / %
WACCM	0.065	0.084	0.050	0.019	-0.015	29.5	-23.4
ULAQ	0.062	0.087	0.049	0.025	-0.013	40.8	-21.1
SOCOLv3	0.109	0.142	0.084	0.033	-0.025	29.9	-23.1
NIWA	0.109	0.134	0.088	0.026	-0.021	23.8	-19.4
MRI	0.090	0.114	0.071	0.024	-0.019	26.6	-21.3
GEOSCCM	0.072	0.089	0.058	0.017	-0.014	24.0	-19.4
EMAC-L90	0.152	0.202	0.114	0.051	-0.038	33.3	-25.2
EMAC-L47	0.181	0.232	0.141	0.051	-0.040	28.2	-22.0
CMAM	0.112	0.145	0.086	0.034	-0.026	30.1	-23.3
CCSRNIES	0.139	0.170	0.113	0.031	-0.025	22.7	-18.3
ACCESS	0.080	0.099	0.064	0.019	-0.016	24.4	-19.7
MMM	**0.106**	**0.136**	**0.083**	**0.030**	**-0.023**	**28.5**	**-21.5**
σ(MMM)	0.036	0.047	0.028	0.011	-0.009	5.0	2.0

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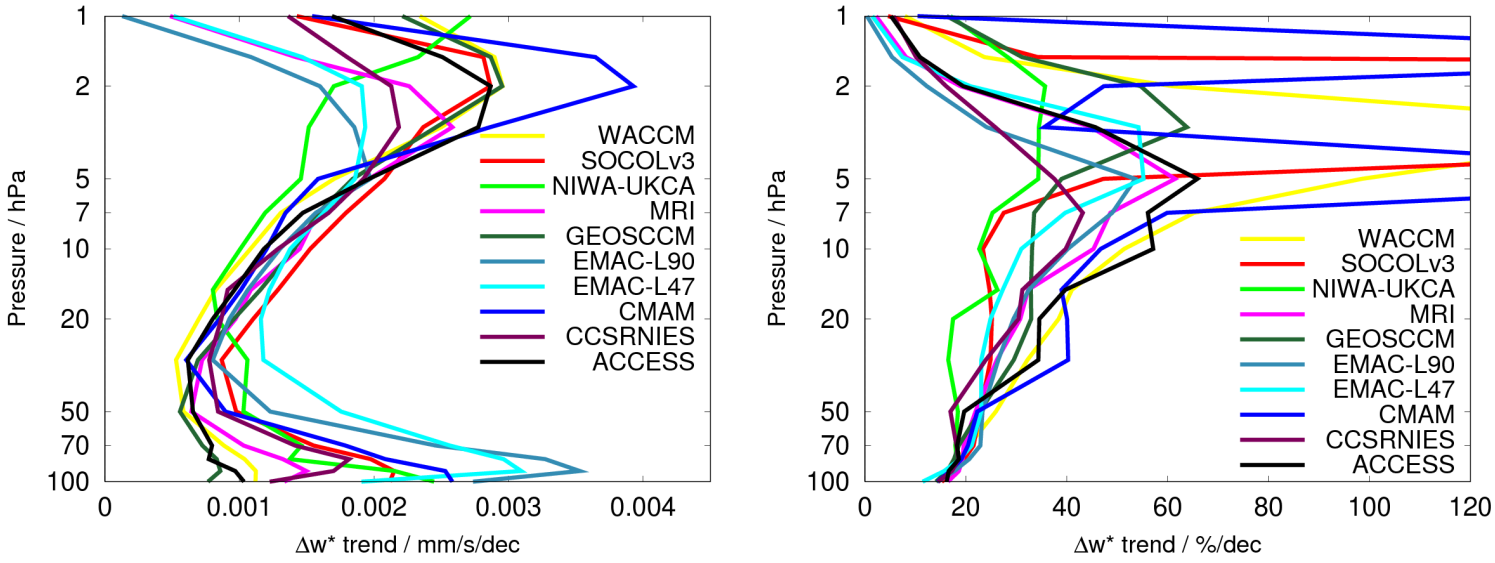
w* trend differences



- Absolute Δw^* has maximum in lower strat., relative Δw^* in mid stratosphere
- Model spread low in mid stratosphere for absolute Δw^*, but growing with altitude for relative Δw^*

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