What controls the meridional distribution of water isotopes?

Nicholas Siler[1], Adriana Bailey[2], Gerard Roe[3], Christo Buizert[1], & David Noone[1]

[1]Oregon State University, [2]NCAR, [3]University of Washington

Background

Single-stream Rayleigh fractionation:

$$R_v = R_e f^{\overline{\alpha}-1}$$
$$R_p = \alpha R_v$$

R_v = Isotope ratio of water vapor
R_p = Isotope ratio of precipitation
R_e = Initial isotope ratio (i.e., at point of evaporation)
f = fraction of initial vapor that has not yet condensed
α = Instantaneous fractionation factor
$\overline{\alpha}$ = Path-integrated α

At a given latitude, R_v and R_p can be represented as the weighted average of all upstream sources:

Time/column averages:

$$\overline{R_v}(x) = \int_{-1}^{1} R_v(x, x^*) w(x, x^*) dx^*$$

$$\overline{R_p}(x) = \alpha(x) \int_{-1}^{1} R_v(x, x^*) w(x, x^*) dx^*$$

Source distribution function:

$$w(x, x^*) = \frac{E(x^*) f(x, x^*)}{\int_{-1}^{1} E(x^*) f(x, x^*) dx^*}$$

x = sin(lat)
$E(x)$ = evaporation
$f(x, x^*)$ = fraction of vapor that **originates at x^*** and is **transported to x** without precipitating

A simple approximation for $w(x, x^*)$

Consider a radiative transfer analogy:

- Vapor transport (F) is "attenuated" by precipitation (P)
  - Attenuation coefficient: $\mu(x) = 2\pi a^2 \frac{P(x)}{|F(x)|}$
  - Optical depth: $\tau(x, x^*) = \int_{x^*}^{x} \mu(\tilde{x}) d\tilde{x}$
  - Transmittance: $f(x, x^*) = \exp[-\tau(x, x^*)]$

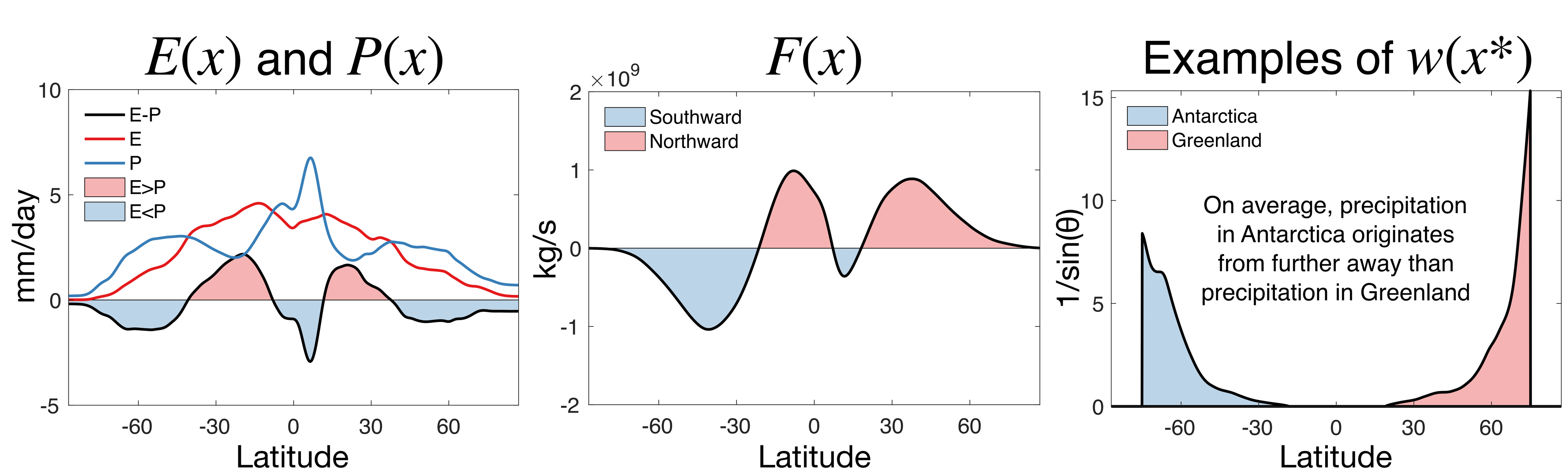
μ^{-1} can be interpreted as a length scale of vapor transport

a = Earth's radius

Vapor transport is set by $E - P$:

$$F(x) = \int_{-1}^{x} [E(x^*) - P(x^*)] dx^*.$$

→ $w(x, x^*)$ can be approximated from $E(x)$ and $P(x)$ *alone.*



Isotope model

In this framework, isotopes are controlled by 3 variables:

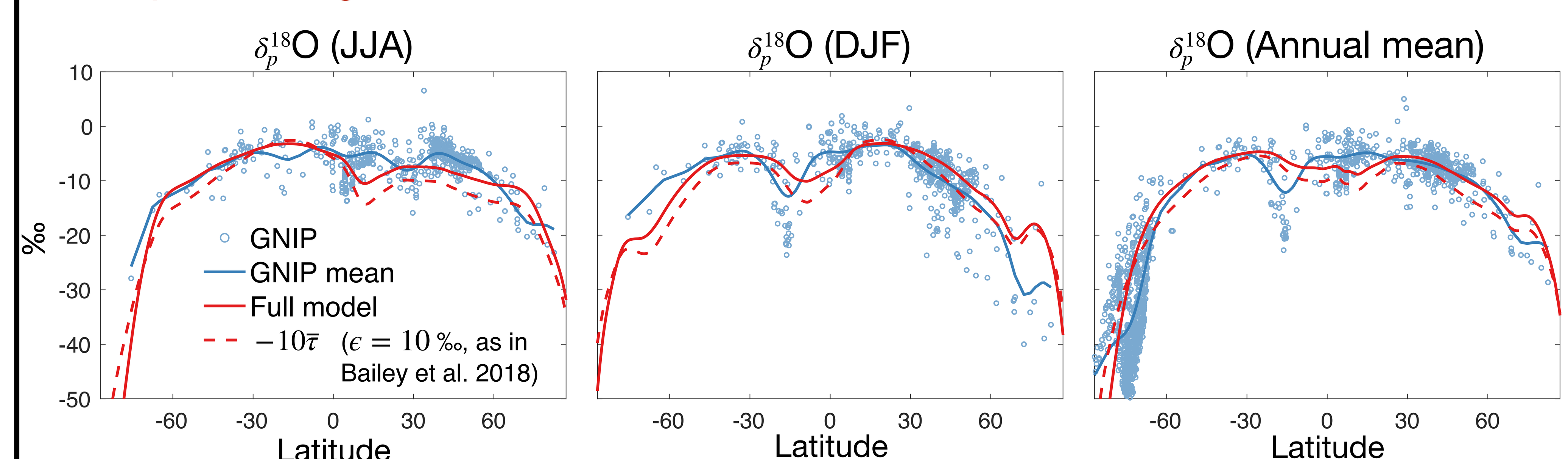
- α = function of temperature at condensation; we parameterize as a function of surface temperature.
- R_e = function of near-surface temperature, relative humidity, & kinetic effects; we use the Craig-Gordon model.
- μ = function of P and F.

If we assume that α and R_e are both constant, then

$$\delta_p(x) \approx -\epsilon \overline{\tau}(x)$$

i.e., δ of precipitation is proportional to the average path-integrated attenuation!

$$\delta_p = \left(\frac{R_p(x)}{R_{std}} - 1 \right) \times 1000$$
$$\epsilon = (\alpha - 1) \times 1000$$
$$\overline{\tau}(x) = \int_{-1}^{1} \tau(x, x^*) w(x, x^*) dx^*$$



Climate change simulations

The 1D Moist Static Energy Balance Model (MEBM)

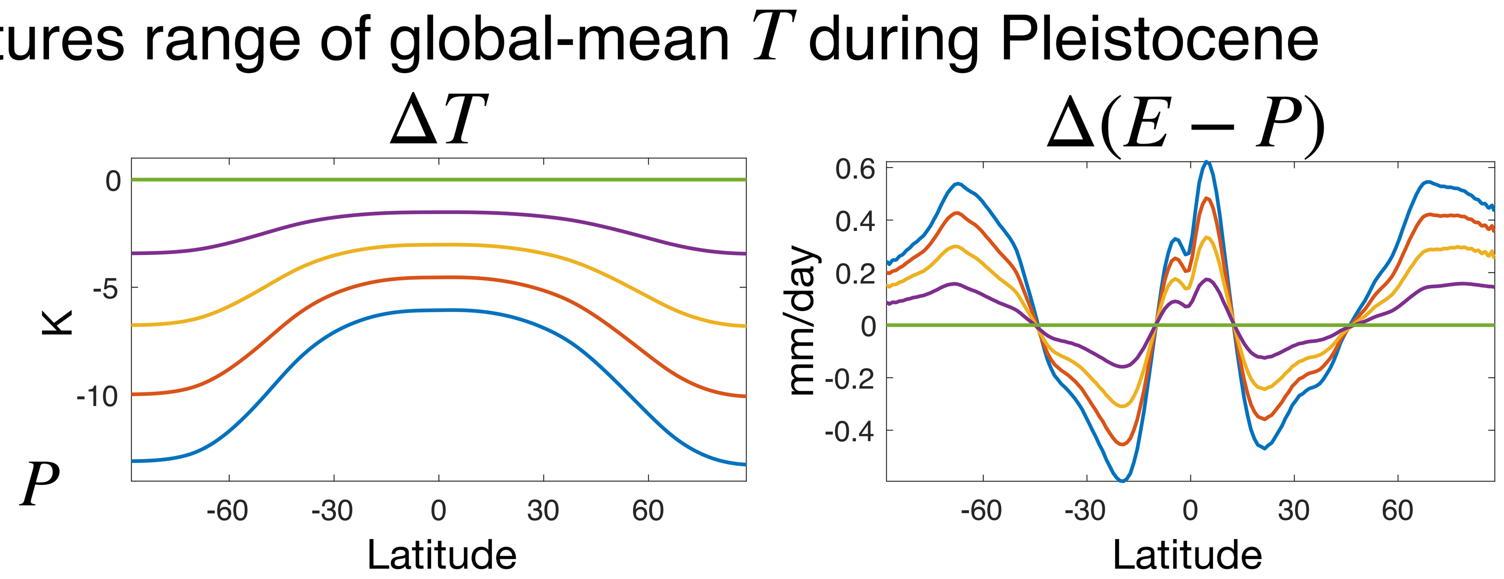
- Down-gradient meridional energy transport
- Hadley Cell parameterization for realistic hydro cycle
- Described in Siler et al. (2018); see also poster by Natalie Burls

Experiment description:

- Impose spatially-uniform radiative feedback of -1 Wm^{-2}K^{-1}
- Impose spatially-uniform TOA forcing ranging from -8 to -2 Wm^{-2}
  - Roughly captures range of global-mean T during Pleistocene

Results:

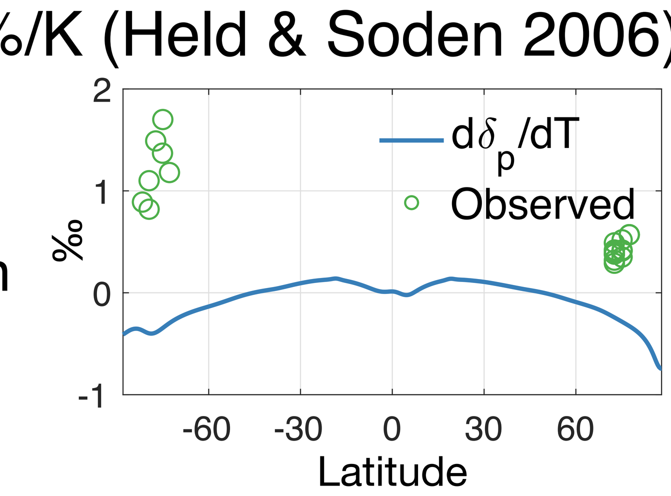
- Polar-amplified cooling
- Amplification of mean-state $E - P$



The isotopic response to climate change

Case 1: E scales uniformly at a rate of ~2%/K (Held & Soden 2006)

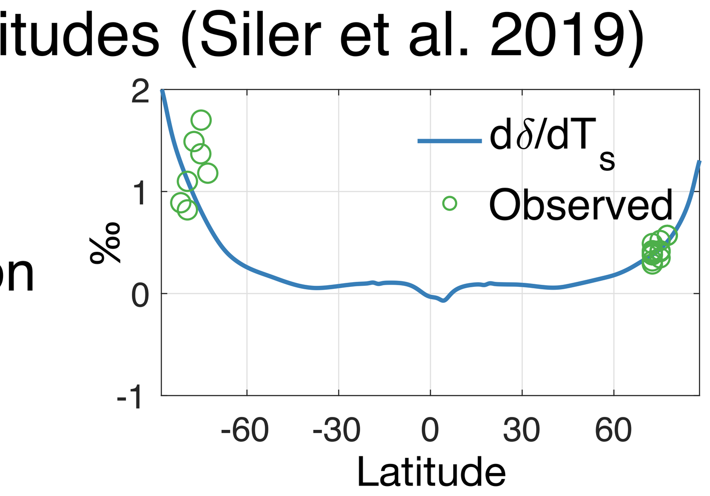
δ_p-vs.-T regression slope



- *Negative* slope at high latitudes
  - μ increases with cooling; P originates from closer by, and is less depleted
- Inconsistent with observations

Case 2: E is more sensitive to ΔT at high latitudes (Siler et al. 2019)

δ_p-vs.-T regression slope



- *Positive* slope at high latitudes
- Suggests that temporal slope primarily reflects thermodynamic changes in spatial pattern of E