LECTURE III

ed gerber

1-D climate models

- Goal: begin to account for the latitudinal structure of the Earths climate.
- before

$$\begin{aligned} \text{energy in} &= \text{energy out} \\ S_0(1-\alpha) &= OLR \\ 0 &= OLR(y) - S_0(1-\alpha) \end{aligned}$$

- now, consider the incoming and outgoing energy as a function of latitude y

$$OLR(y) - S_0(y)(1-\alpha(y)) = F(y)$$

1-D climate models

- F(y) represents a local imbalance between incoming and outgoing energy

$$\int_{SP}^{NP} dy\, OLR(y) - S_0(y)(1-\alpha(y)) = \int_{SP}^{NP} dy\, F(y)$$

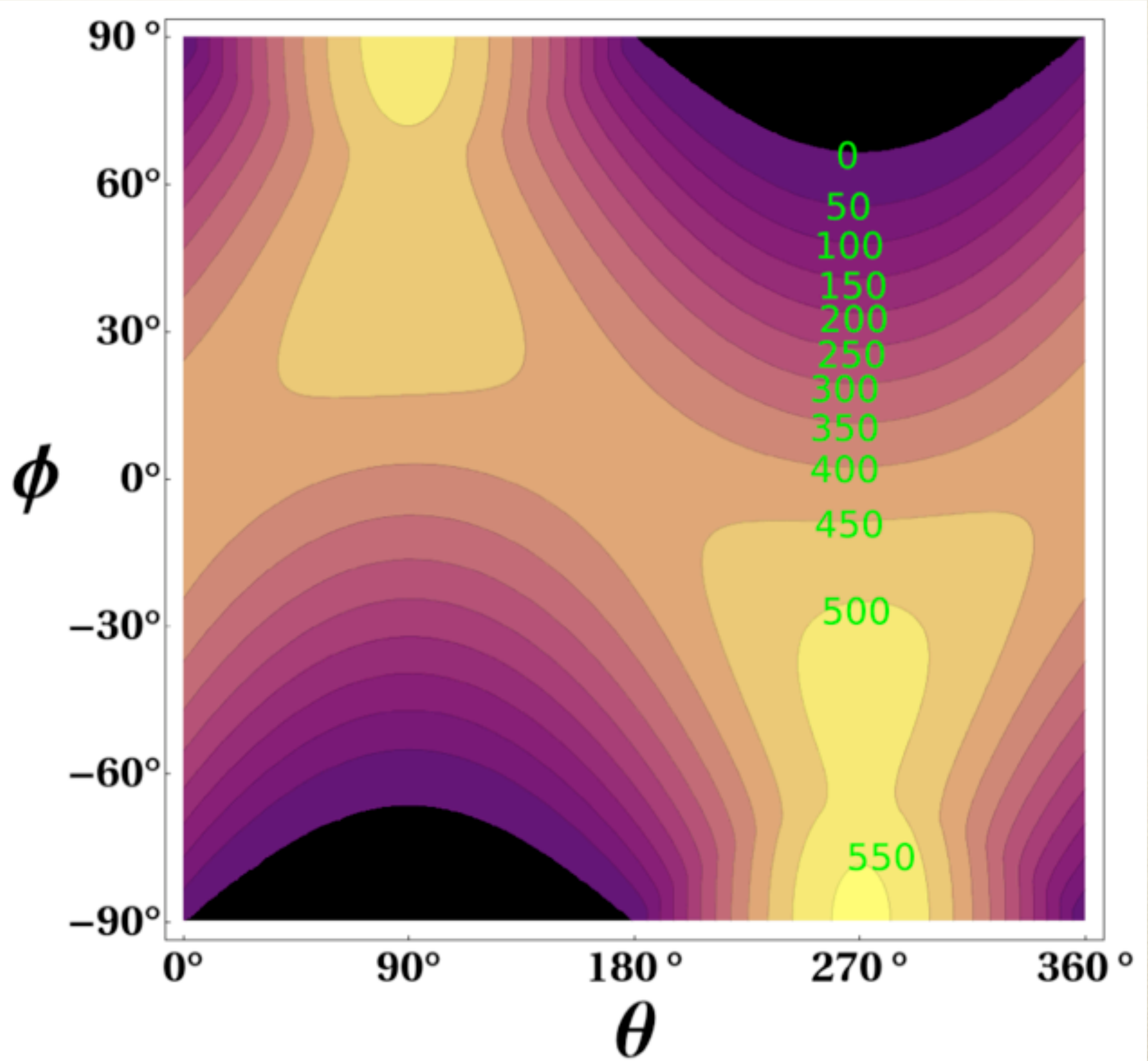
$$0 = \int_{SP}^{NP} dy\, \mathrm{F}(y)$$

- Think of F(y) as the impact of sensible and latent heat transport! You can't add or subtract net energy, but you can move it from one latitude to another.

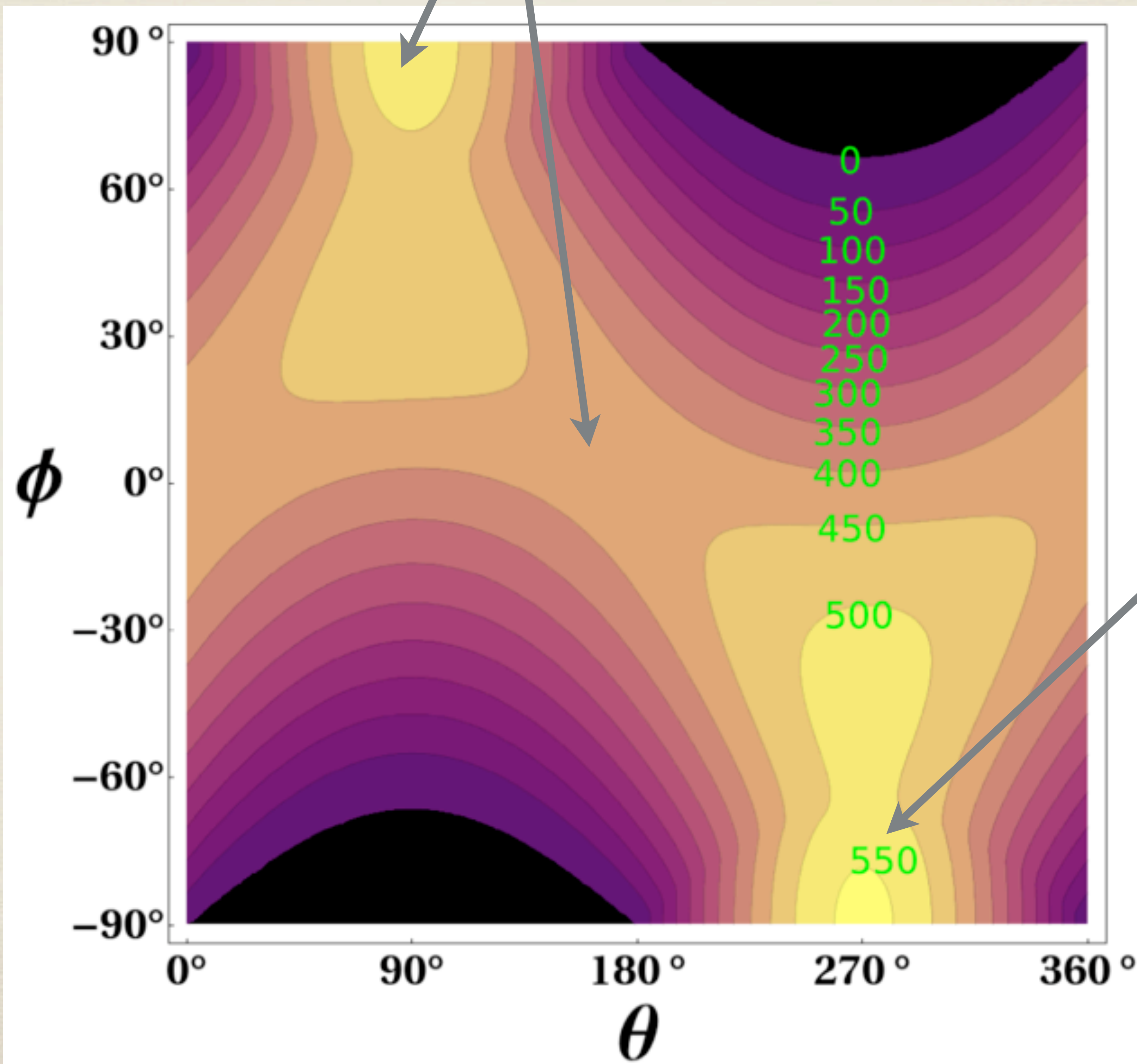
Insolation (latitude)

- Insolation as a function of latitude, So(y)= So f(y,t), where f(y,t) is a "flux factor" which depends on latitude (y) and time of the year (t).
- Two competing factors
  - inclination angle (more net radiation when sun is overhead, as in tropics)
  - length of day (more sunshine = more radiation)
- Tilt (obliquity) of the earth key to seasonal dependence.

Insolation So(y,t)



poles gets more sun in summer than equator ever gets!

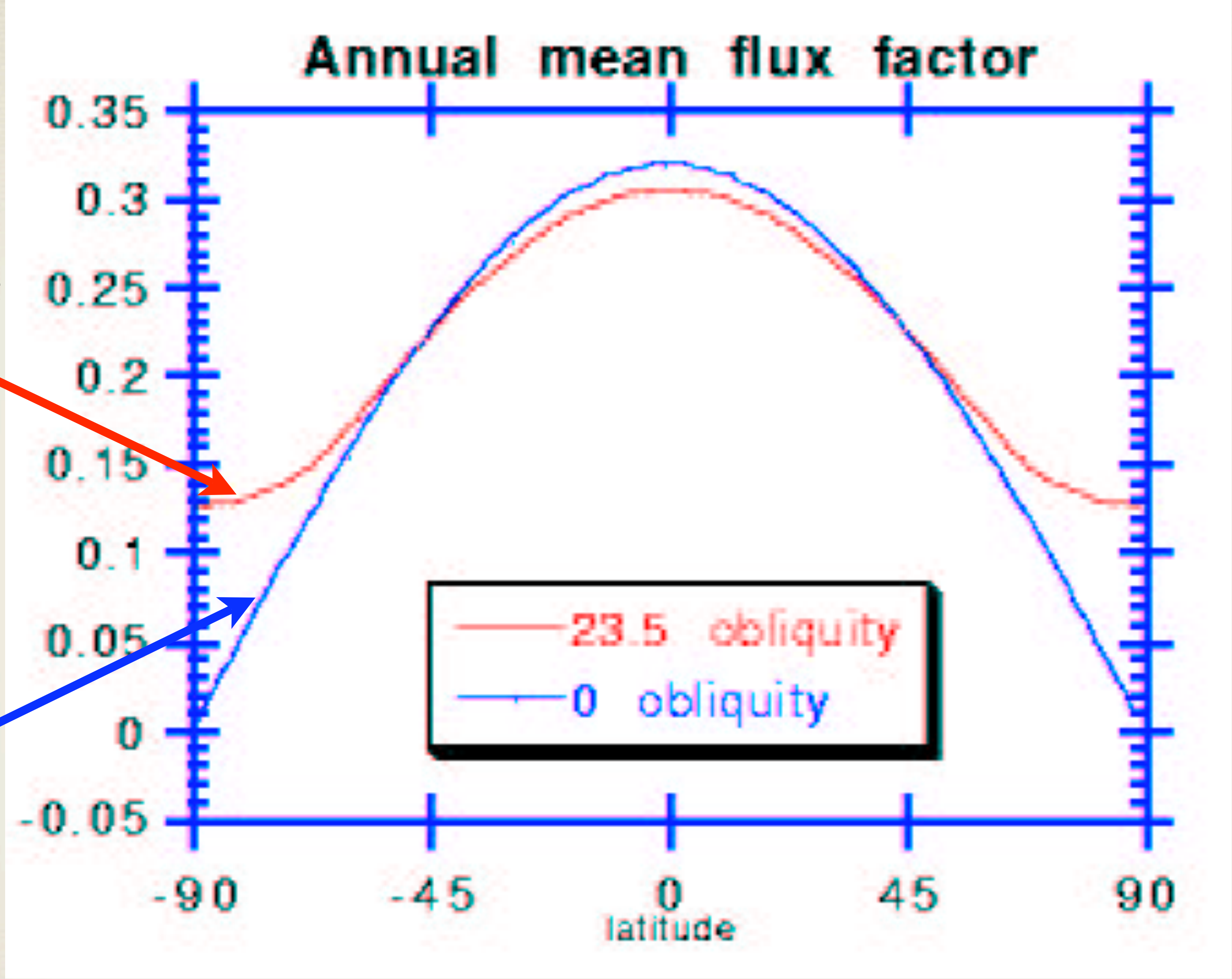


SH summer gets a local max insolation because earth orbit is closer during this period of the year!

- Experience tells us that the south pole in January (austral summer) is not the warmest place to be!
- Earths climate averages out radiation! (albedo...)

Annual mean solar radiation is a more meaningful metric for simple climate models.

w/ tilt,
cold pole



w/o tilt
COLD pole

Earth's obliquity (tilt) has a huge impact on poles. (Without tilt, you'd only have the inclination effect.)
[hmm, change the tilt, change the climate!]

Some back of the envelope calculations...

- Even with obliquity, poles receive less than half as much sunlight as equator. On top of that, snow reflects with average albedo 0.7, ice free albedo averages 0.1.
- Recall our "zeroth order mode" for temperature

$$T_g = \sqrt[4]{\frac{S_0 f(y)(1-\alpha)}{\sigma(1-\varepsilon/2)}}$$

- So/4 becomes So*f(y), as we're accounting for the local averaging of inclination + length of day.

Some back of the envelope calculations...

- At the equator:

$$T_g = \sqrt[4]{\frac{1367 \cdot 0.3(1-0.1)}{\sigma(1-\varepsilon/2)}} = 320$$

320 K = 47 C = 116 F (bit warm)

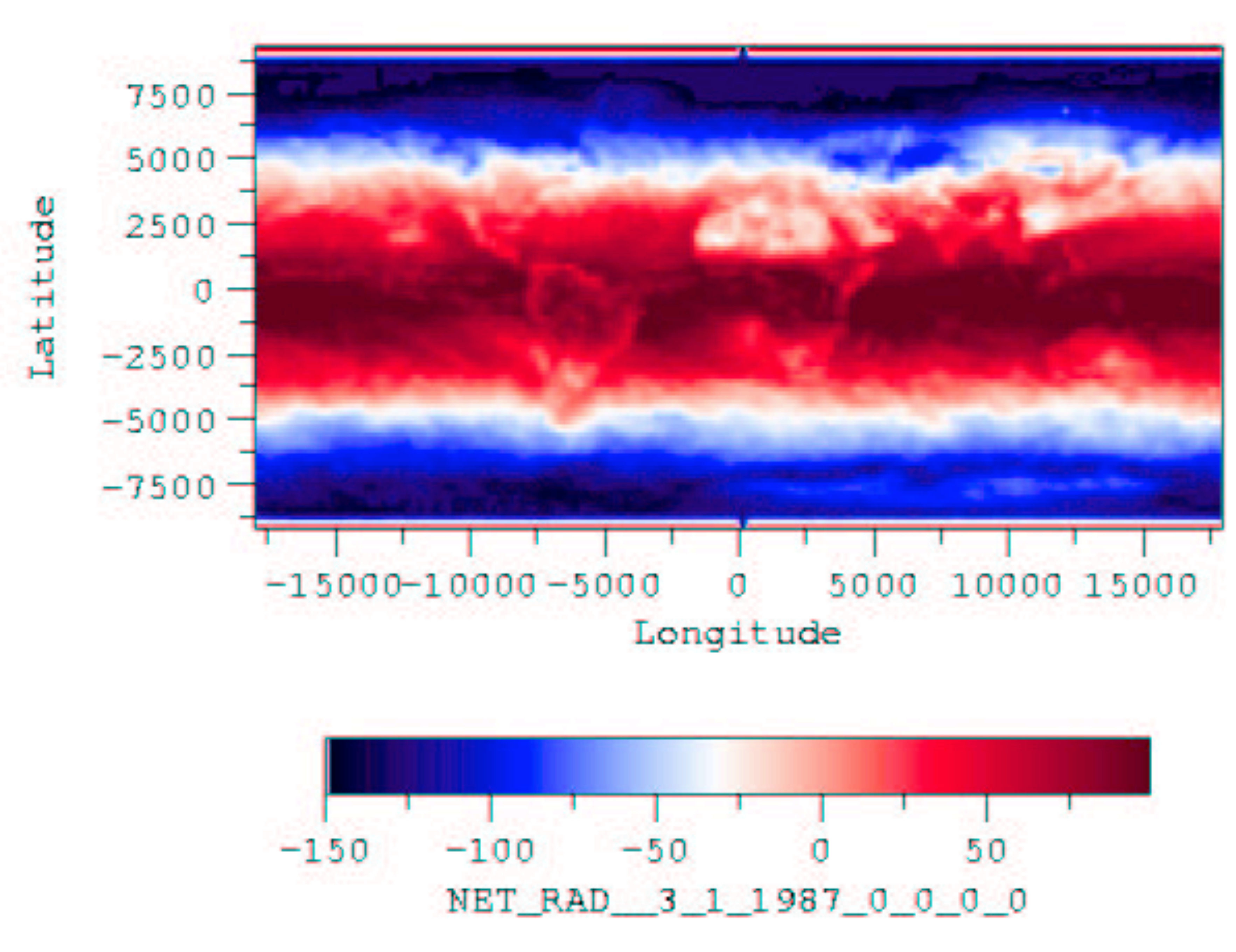
- At the pole:

$$T_g = \sqrt[4]{\frac{1370 \cdot 0.13(1-0.7)}{\sigma(1-\varepsilon/2)}} = 197$$

197 K = -78 C = -105 F (CO_2 at 1 atm freezes at -78.5 C)

- Real situation worse, as emissivity depends on T, as water vapor the top greenhouse gas!

Radiative Imbalance

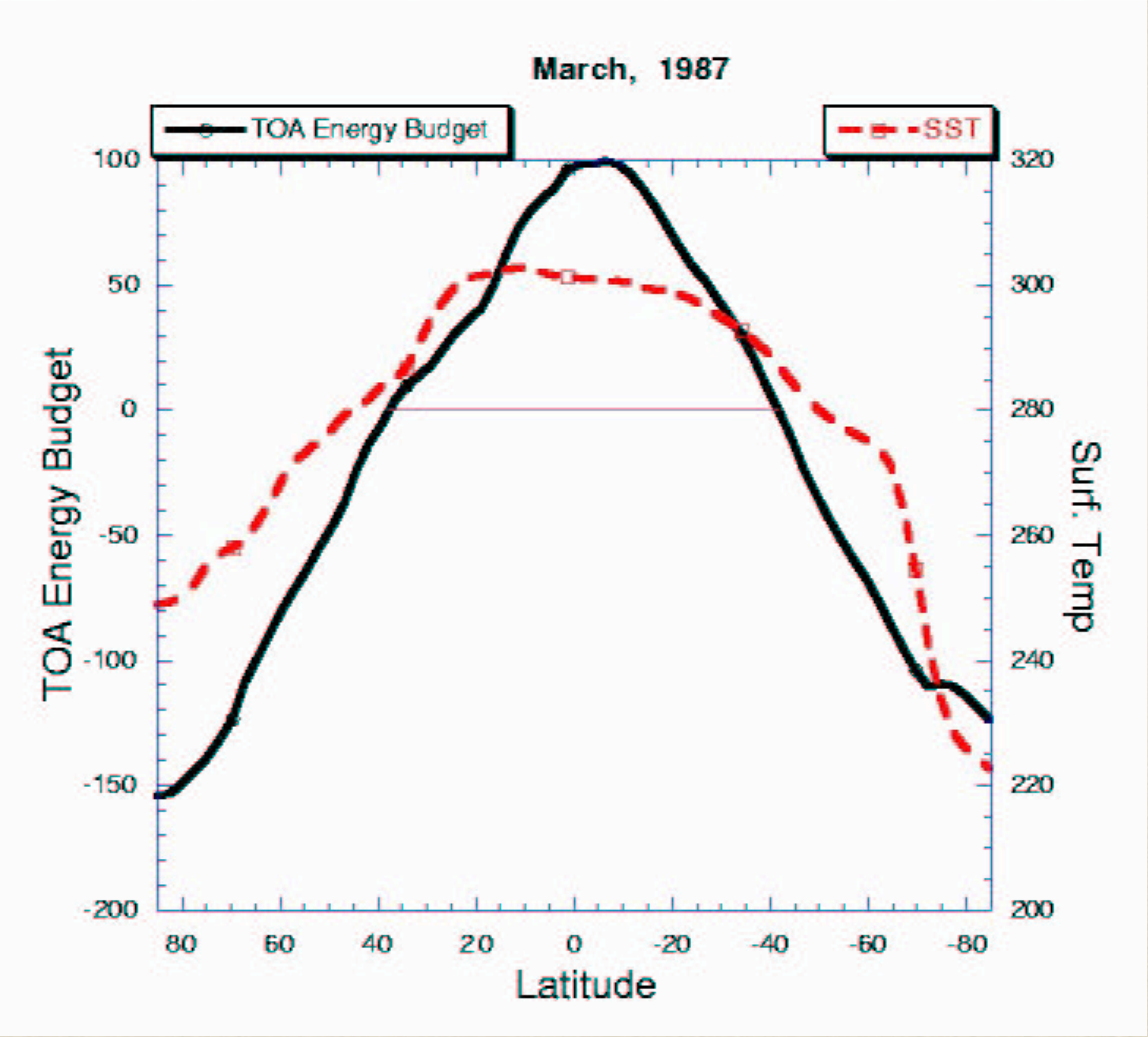


red = more energy in than OLR out

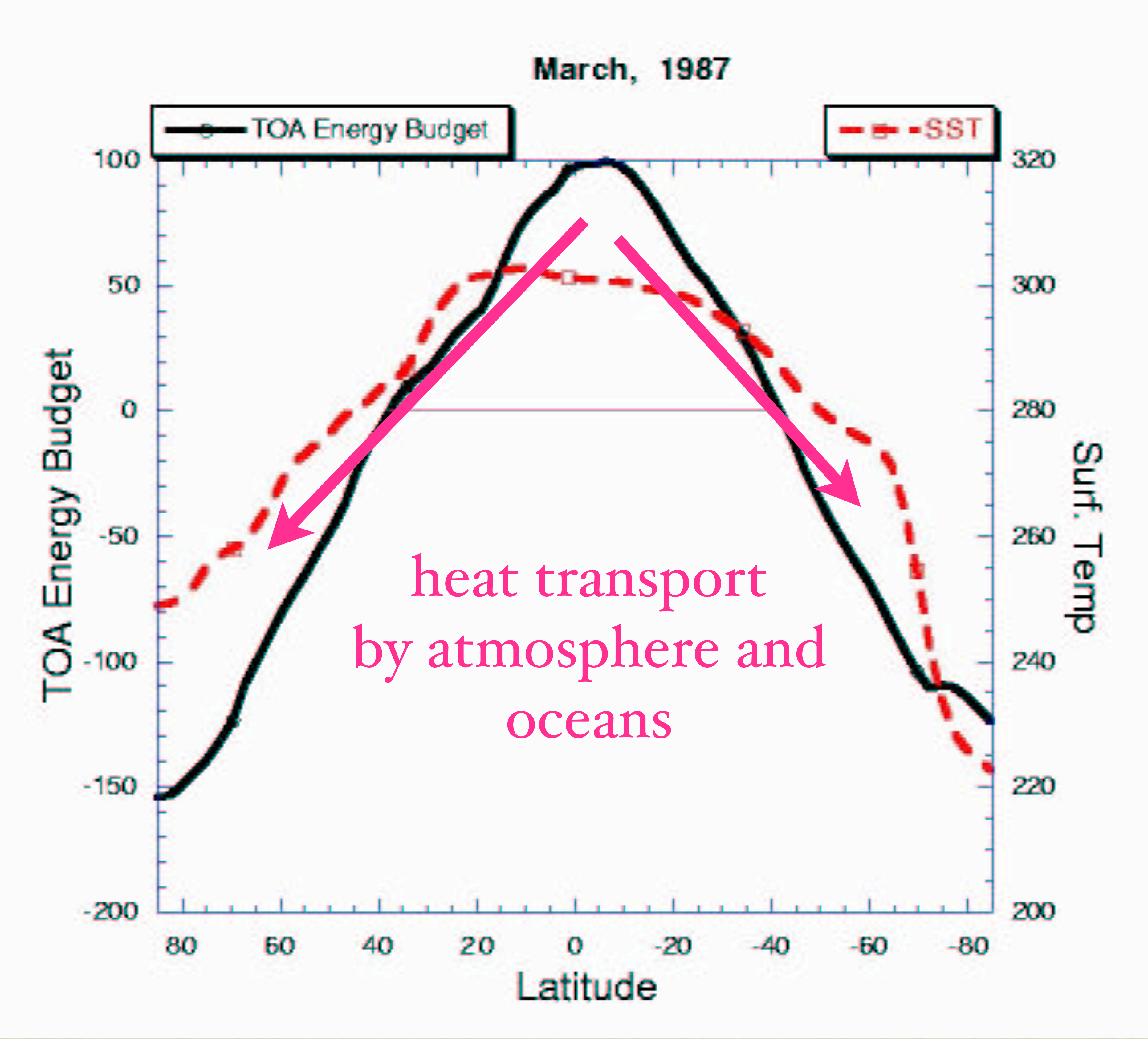
blue = less energy in than OLR out

(note albedo of Sahara desert, less energy in there, blackness=strong absorption in Amazon and Congo rainforest)

Zonal Mean Imbalance



Zonal Mean Imbalance



Atmosphere + Ocean = heat transport

* we've discussed the key uncertainties associated with aerosols and clouds on both the emissivity and albedo, which require modeling of the atmosphere + ocean
* another key need to model dynamics of atmosphere + oceans are to understand the transport of heat
* Last of our energy balance models: model atmosphere and ocean as "diffusive processes" which diffuse heat.
* Sellers (J. Applied Meteorology, 1969) and Budyko (Tellus, 1969)

[See links from course page for these historic papers!]

A Diffuse Energy Balance Model

Consider local coordinate system: y = sin(latitude).

Suppose temperature is diffused:

$$F(y) = \frac{d}{dy}\left((1-y^2)D\frac{dT}{dy}\right)$$

Suppose hemispheres are symmetric. Then boundary conditions are $T_y=0$ at y=0 (no flux across equator) and T(1)=T(pole) must be regular.

A Diffuse Energy Balance Model

Climate equation becomes:

$$
\begin{aligned}
F(y) &= OLR(y) - S_0(y)(1-\alpha(y)) \\
\frac{d}{dy}\left((1-y^2)D\frac{dT}{dy}\right) &= OLR(y) - S_0 f(y)(1-\alpha(y))
\end{aligned}
$$

Linearize the OLR about T*, a reference climate temperature.

$$OLR(y) = B(T - T^*) = BT'$$

to get

$$\frac{d}{dy}\left((1-y^2)\frac{dT'}{dy}\right) = \frac{B}{D}T' - \frac{S_0}{D}f(y)(1-\alpha(y))$$

A Diffuse Energy Balance Model

Lastly, we'll make a simple albedo assumption. Let y_i be the "snow line." Polewards, earth is ice covered.

$$\alpha = \begin{cases} \alpha_0 \approx 0.1, & y < y_i \\ \alpha_i \approx 0.7, & y \geq y_i \end{cases}$$

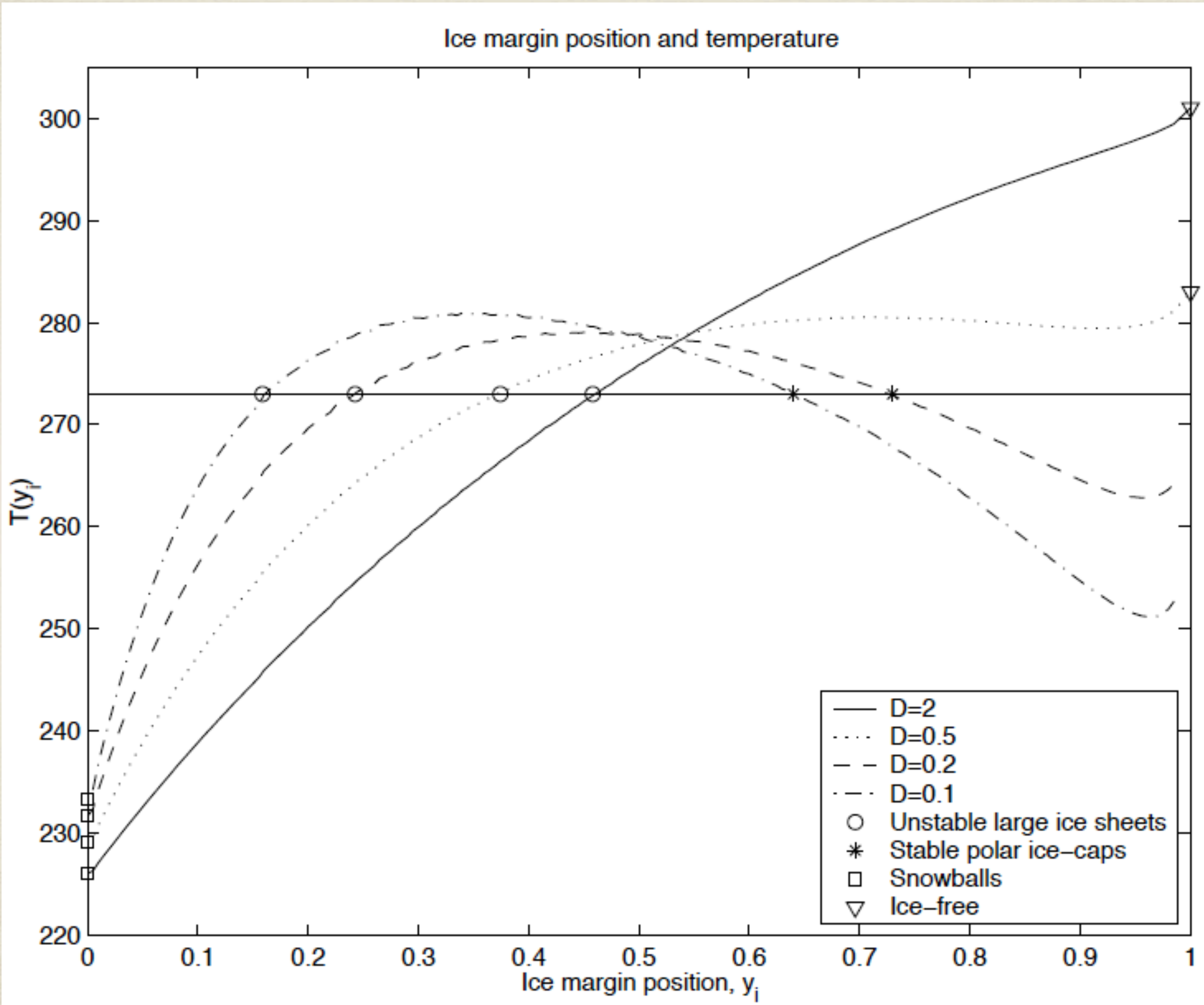
Goal is to adaptively fit y_i to the resulting temperature profile so that $T(y_i)=273$: this gives you a self consistent climate!

$$\frac{d}{dy}\left((1-y^2)\frac{dT'}{dy}\right) = \frac{B}{D}T' - \frac{S_0}{D}f(y)(1-\alpha(y))$$

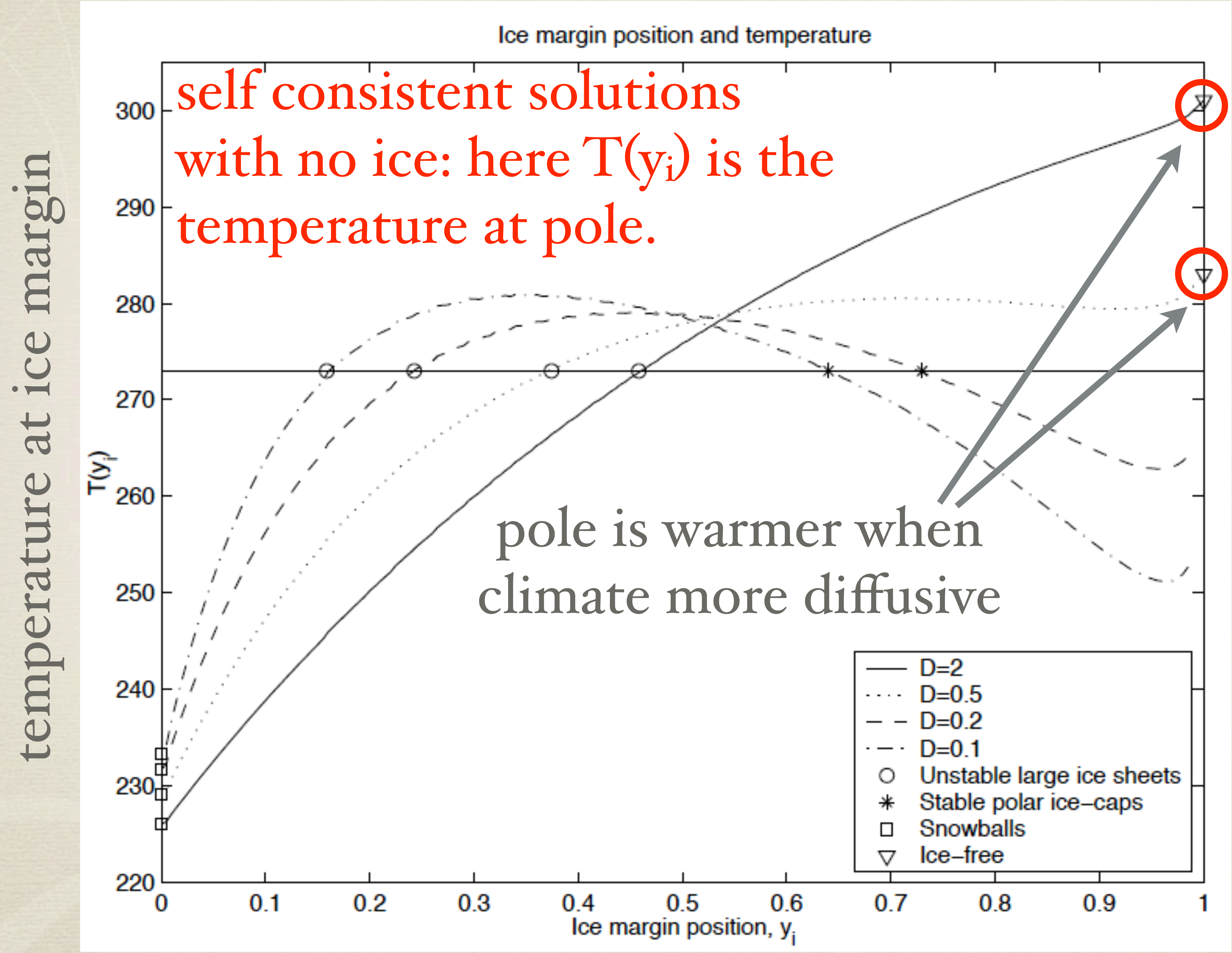
(remember that f(y) is given function)

The Results

temperature at ice margin

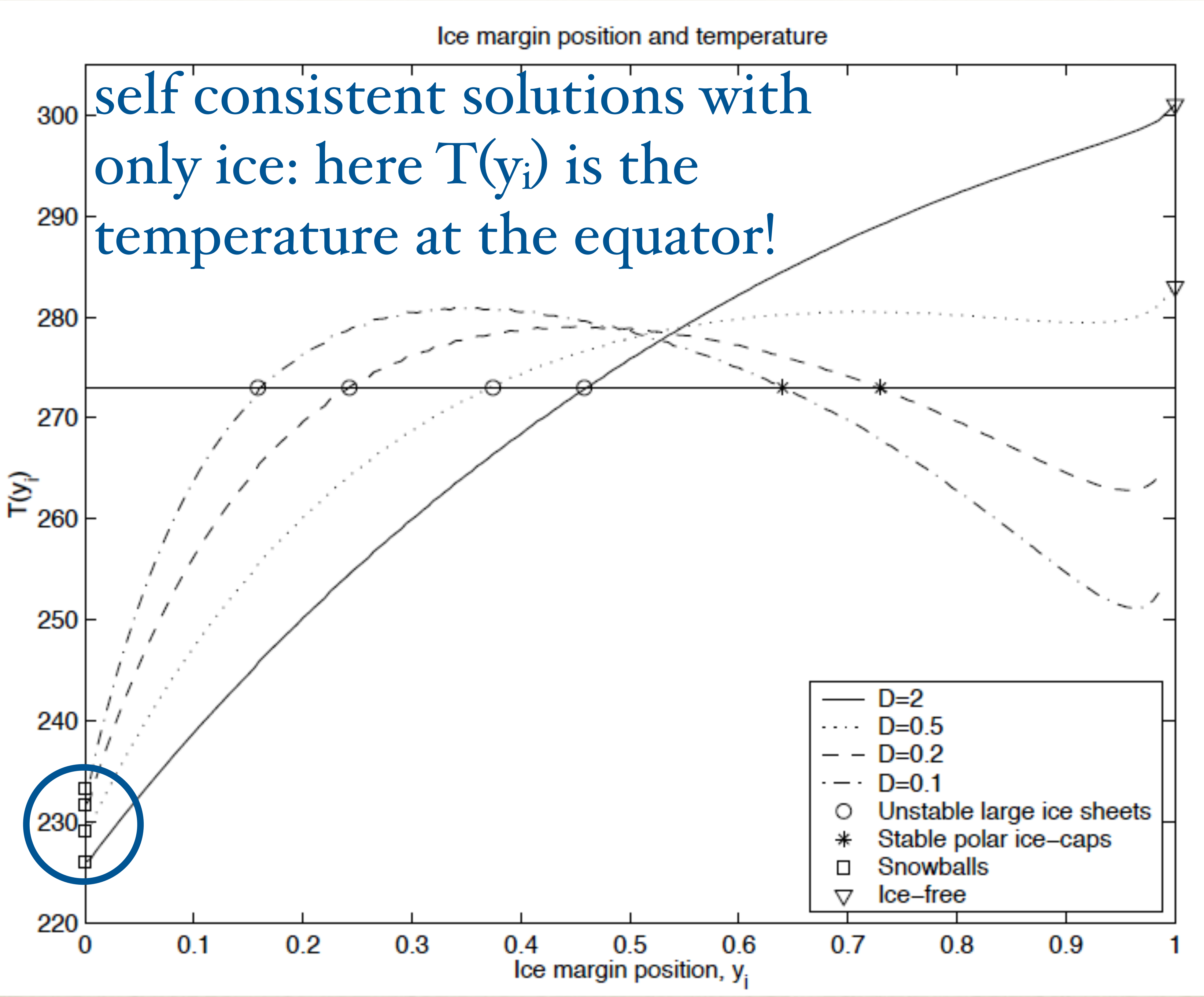


The Results


temperature at ice margin
Ice margin position and temperature
self consistent solutions with no ice: here $T(y_i)$ is the temperature at pole.
pole is warmer when climate more diffusive
$T(y_i)$
Ice margin position, y_i
300
290
280
270
260
250
240
230
220
0
0.1
0.2
0.3
0.4
0.5
0.6
0.7
0.8
0.9
1
D=2
D=0.5
D=0.2
D=0.1
Unstable large ice sheets
Stable polar ice–caps
Snowballs
Ice–free

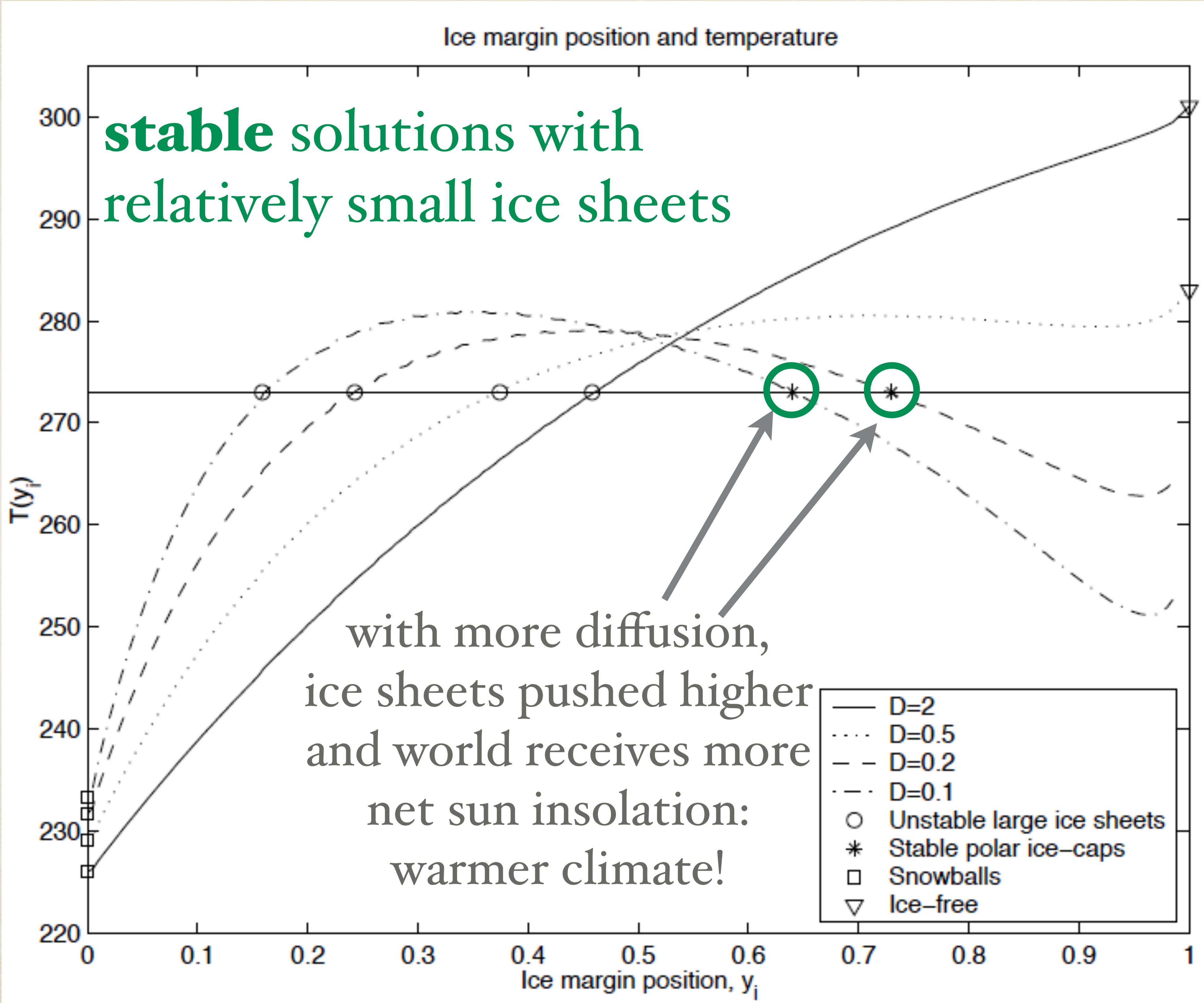
The Results

temperature at ice margin

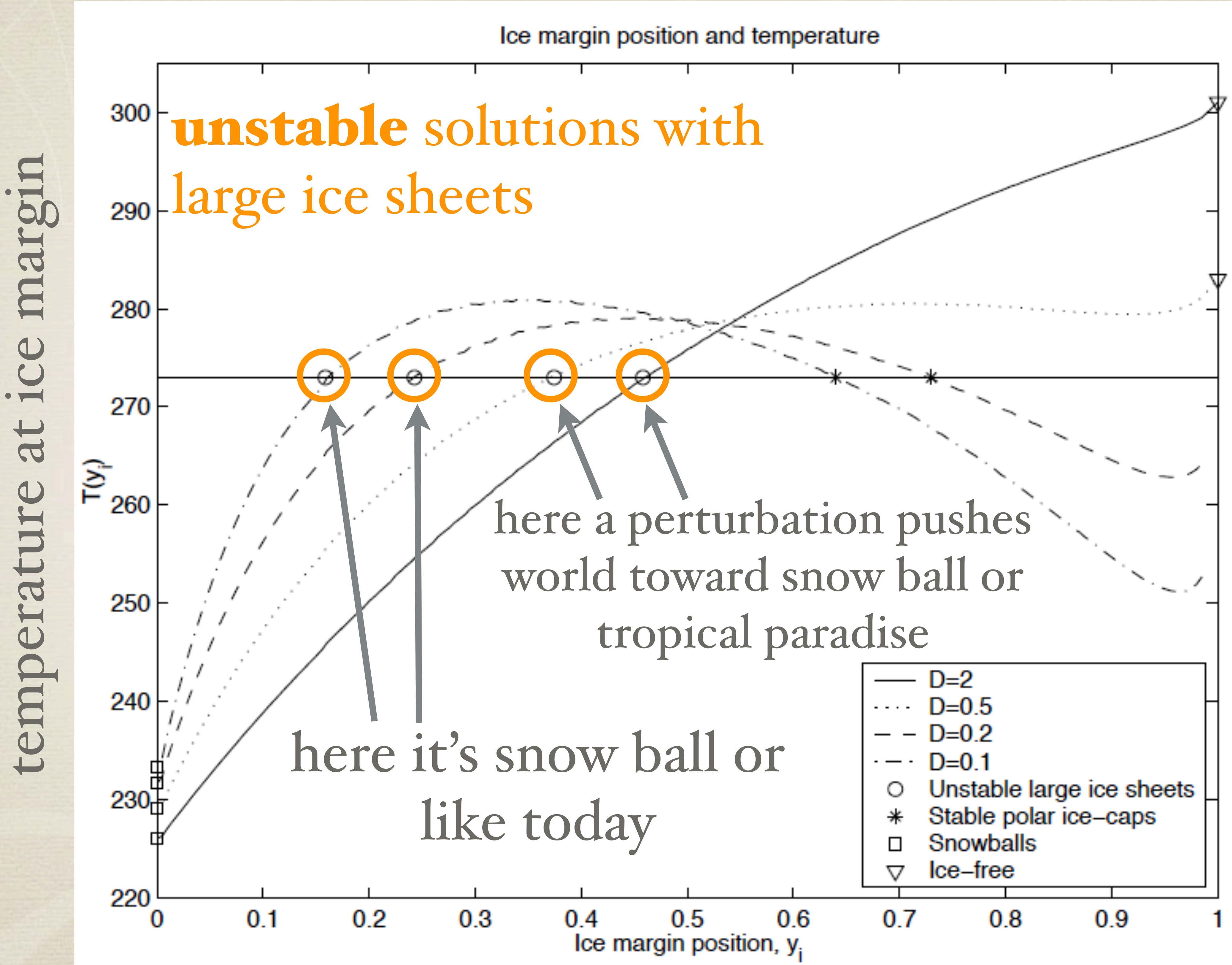


The Results

temperature at ice margin



The Results

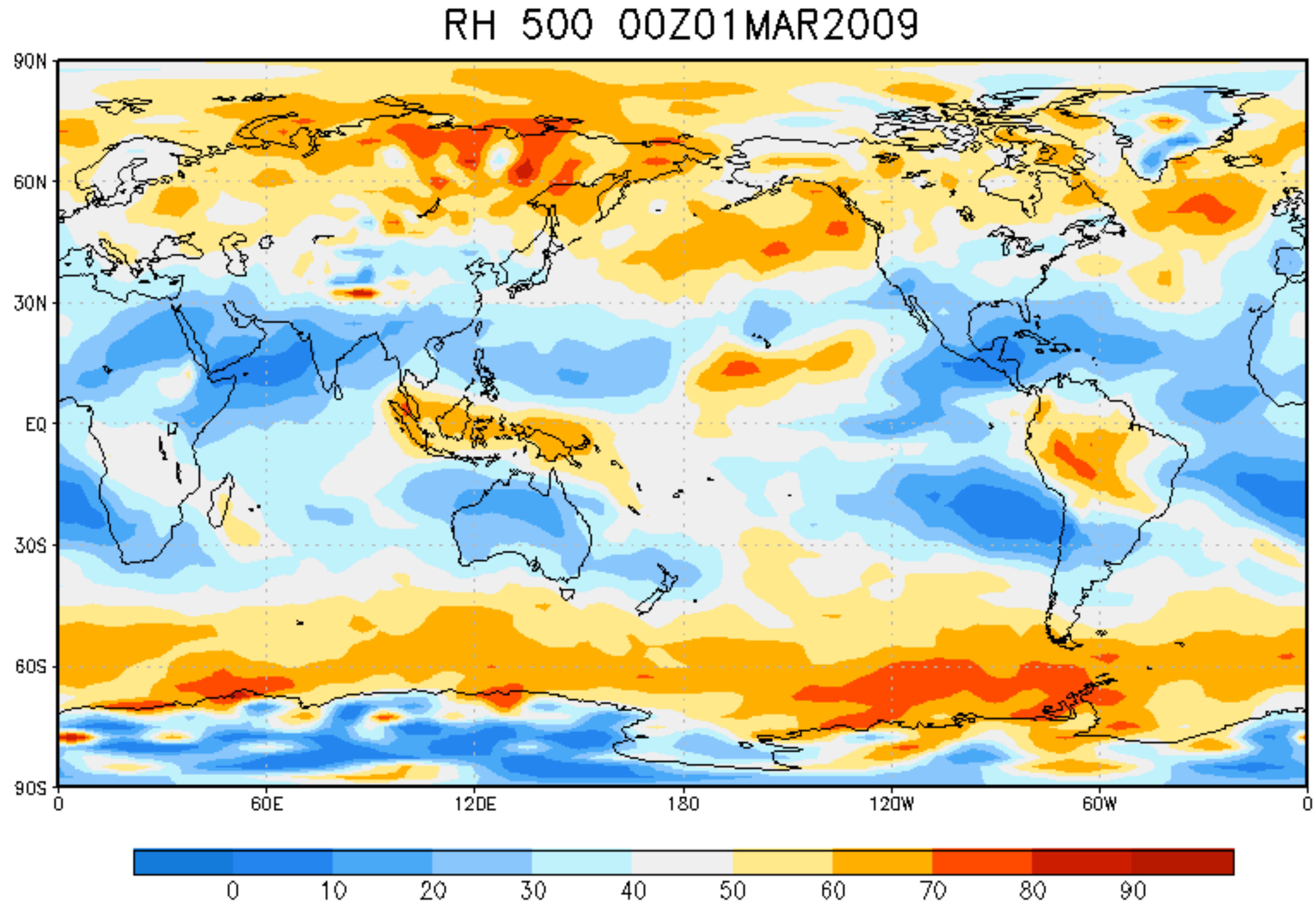


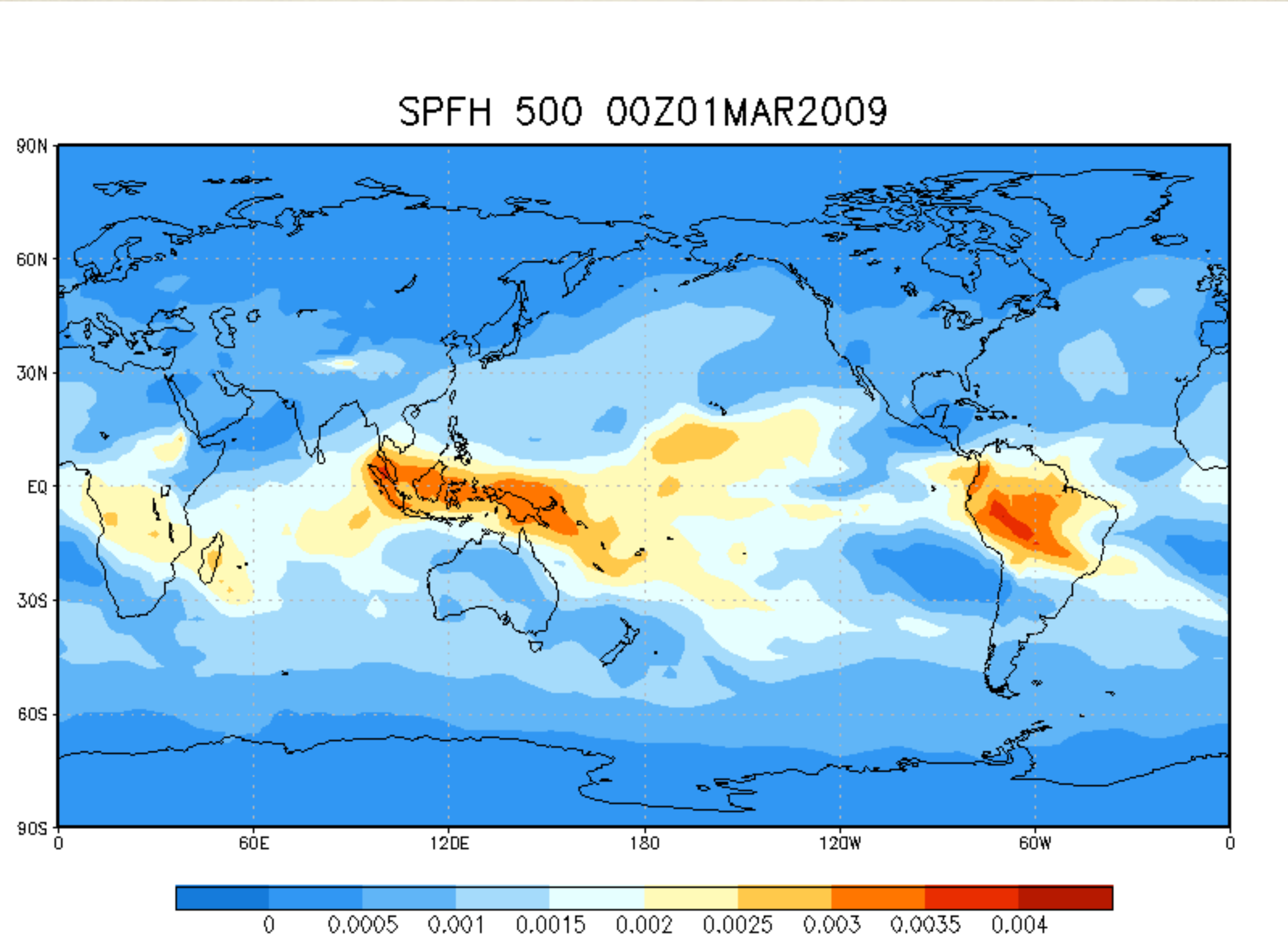
temperature at ice margin
Ice margin position and temperature
unstable solutions with
large ice sheets
here a perturbation pushes
world toward snow ball or
tropical paradise
here it's snow ball or
like today
T(y_i)
Ice margin position, y_i
300
290
280
270
260
250
240
230
220
0
0.1
0.2
0.3
0.4
0.5
0.6
0.7
0.8
0.9
1
D=2
D=0.5
D=0.2
D=0.1
Unstable large ice sheets
Stable polar ice–caps
Snowballs
Ice–free

Final Comments

- $(D/B)^{1/2}$ sets a natural length scale for diffusion. If it's large, you end up with a zeroth order climate model again.
- Linearization of the OLR curve has it's limits. No runaway greenhouse possible here. We could use the full OLR curve for numeric solutions, but then again, this model's really simple! (Law of diminishing returns!)
- One can keep time dependence, $M(dT/dt)$, incorporating thermal inertia of the climate system. This is important for climate transitions, as in the Sellers and Budyko models!

Preview of things to come




SPFH 500 00Z01MAR2009
90N
60N
30N
EQ
30S
60S
90S
0
60E
120E
180
120W
60W
0
0
0.0005
0.001
0.0015
0.002
0.0025
0.003
0.0035
0.004