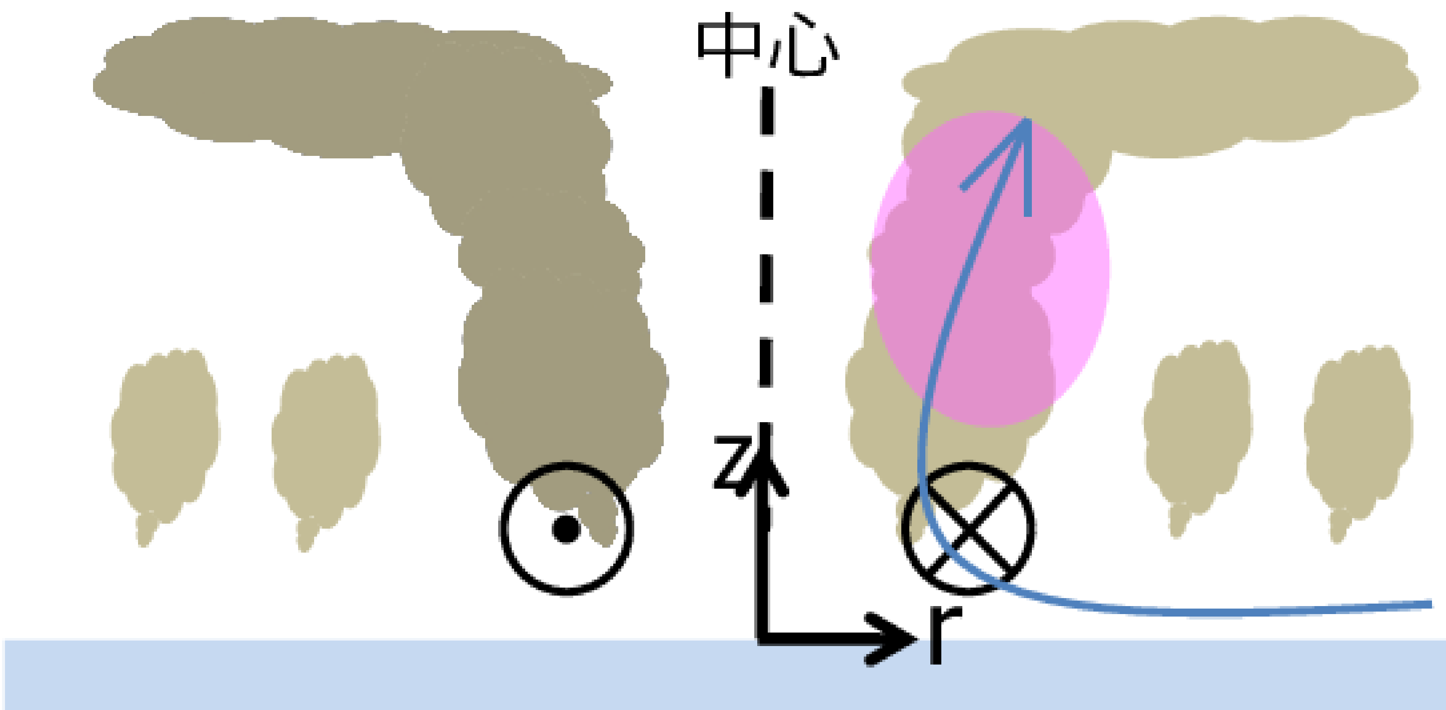
Part 2 WISHEメカニズム

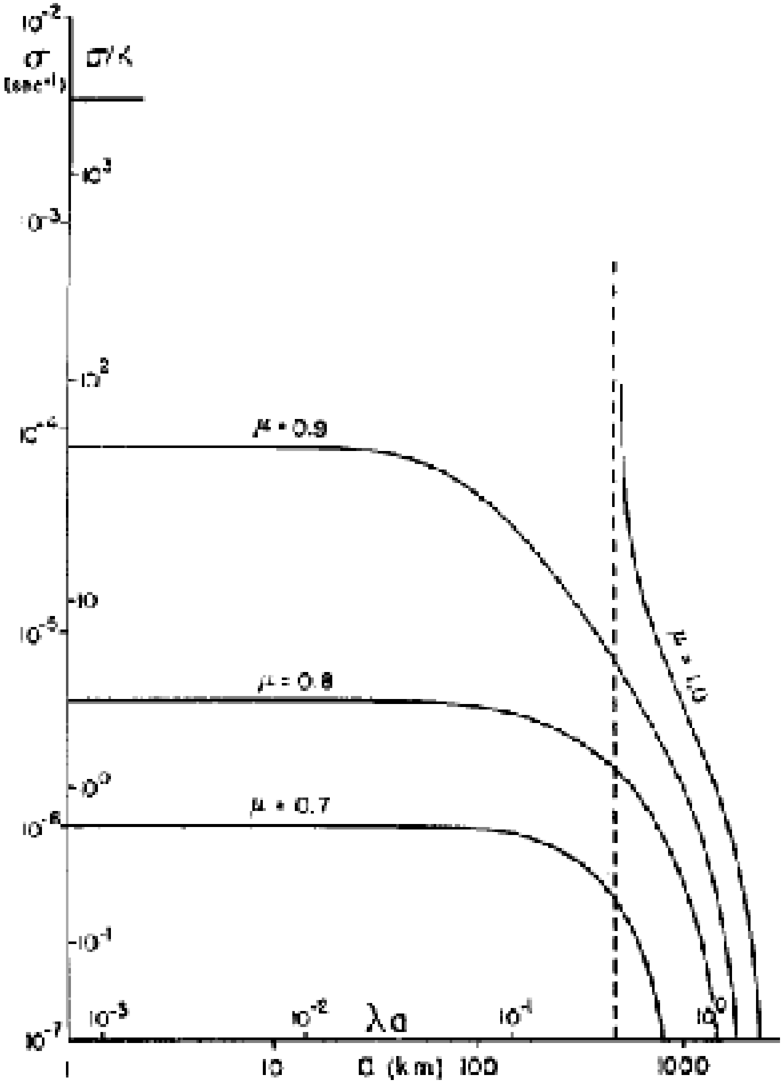
CISK
(Charney & Eliassen, 1964など)

- CISK: Conditional Instability of the Second Kind
  - 対流に伴う凝結熱
    →湿った空気の台風スケールの摩擦収束
    →対流活動の活発化
- 線形擾乱の成長率を計算



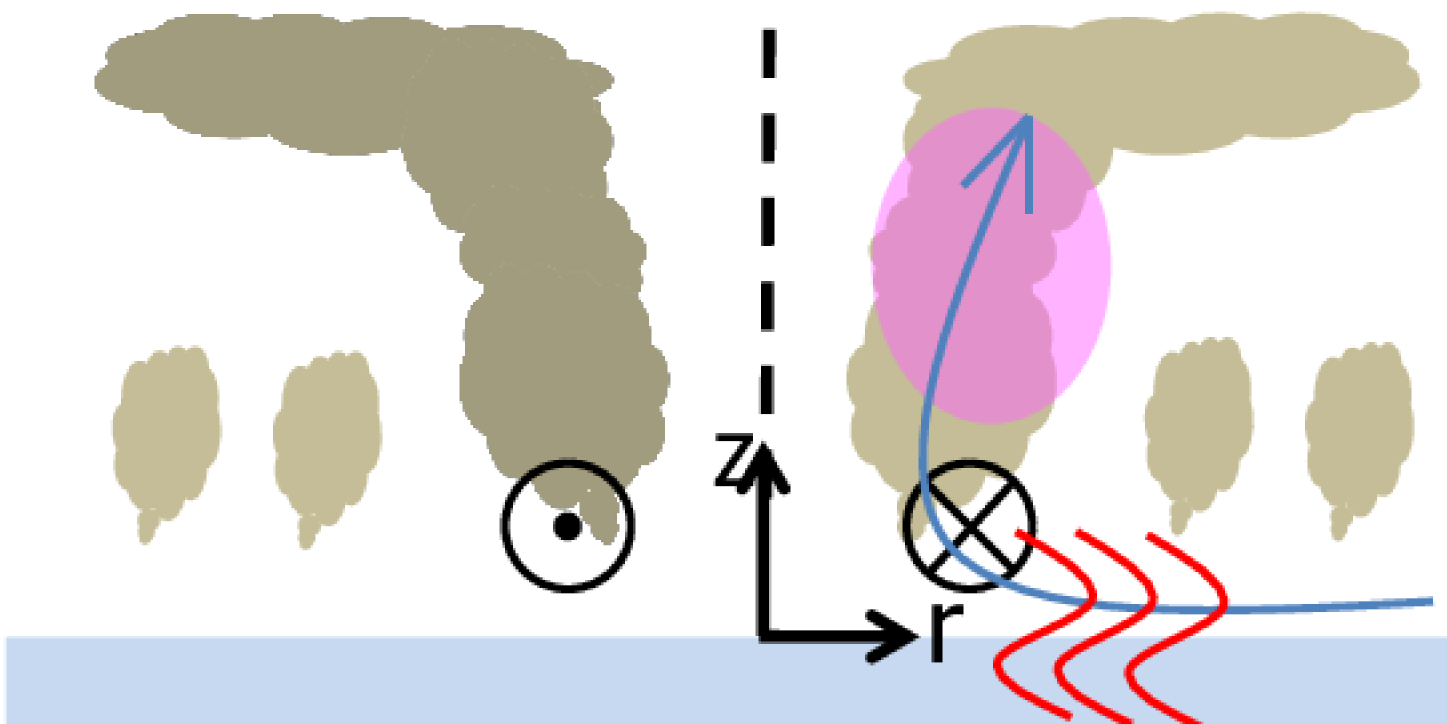
Charney and Eliassen (1964)のCISKを台風の発達に用いる問題点

- 成長率が台風スケールよりも短い波長で大きい値になる(これはOoyama 1969の線形モデルも同様)
- 非断熱加熱が安定度を変える効果が含まれていない。
- より詳細な議論はSmith(1997)を参照



WISHE (Emanuel, 1986)

- WISHE: Wind Induced Surface Heat Exchange
  - ➢ 強風に伴う海面からの水蒸気及び顕熱供給が台風のエネルギーバランスにとって本質的と考えた
  - ➢ 台風をカルノーサイクルに近いものとして説明
  - ➢ <u>定常状態の台風強度の解析解を与えた(のちの数値計算等による時間発展問題もWISHEと呼んでいる)</u>




目が半開きの伊藤
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WISHE

Yano and Emanuel (1991) coined the term WISHE (wind-induced surface heat exchange) to denote the source of fluctuations in subcloud-layer entropy arising from fluctuations in surface wind speed. This can be an important source of energy for tropical disturbances (Emanuel 1987; Neelin *et al.* 1987; and Numaguti and Hayashi 1991. The latter two used the term 'evaporation–wind feedback' to characterize the same process).

(Emanuel et al., 1994)

The Wind-Induced Surface Heat Exchange (WISHE) Model was originally proposed by Emanuel (1987; hereinafter E87) and independently by Neelin et al. (1987) as an alternative to the wave–CISK mechanism for explaining the 30–60-day oscillation. The acronym WISHE is intended to replace and unify the terms "air–sea interaction" used by E87 and "evaporation–wind feedback" used by Neelin et al. (1987).

(Yano and Emanuel, 1991)

Emanuel (1986), echoing the earlier work by Riehl and Kleinschmidt, proposed instead that "the intensification and maintenance of tropical cyclones depend *exclusively*[1] on self-induced heat transfer from the ocean," arguing that ambient conditional instability plays essentially no role, with energy supplied exclusively by surface enthalpy fluxes. A key adjective in this formulation is "self-induced," the idea being that the winds associated with the tropical cyclones drive the surface enthalpy fluxes that power it—a process that has since been called "wind-induced surface heat exchange" (WISHE).

(Zhang and Emanuel, 2016)

CISK

It is proposed that the cyclone develops by a kind of secondary instability in which existing cumulus convection is augmented in regions of low-level horizontal convergence and quenched in regions of low-level divergence. The cumulus and cyclone-scale motions are thus to be regarded as cooperating rather than as competing—the clouds supplying latent heat energy to the cyclone, and the cyclone supplying the fuel, in the form of moisture, to the clouds.

(Charney and Eliassen, 1964)

The model assumes that the large-scale hydrodynamical aspects of a tropical cyclone may be represented by an axisymmetric, quasi-balanced vortex in a stably stratified incompressible fluid, while the effects of moist convection may be formulated through the first law of thermodynamics applied to an implicit model of penetrative convective clouds. The air-sea exchange of angular momentum as well as latent and sensible heat is explicitly calculated in the model with the use of conventional approximations.

(Ooyama, 1969)

The present author views CISK in terms of the conceptual content that has grown and matured with advances in modeling work. Then, the spirit of CISK as the co-operative intensification theory is valid and alive. It is unfortunate, however, such a view of CISK does not seem to be shared by the majority of users of the acronym.

(Ooyama, 1982)

←大山先生はCharney and Eliassen (1964)やOoyama(1969)で考えた線形モデルでは、スケールの問題を解決できないと考えた。CISKの精神は数値計算の世界に生きている

伊藤の理解する”CISK”と”WISHE”の関係

- 人によって定義が違う。
  - 台風研究者A
    - 大気環境場の不安定に起因するのがCISK、大気環境場を中立とみなし海面過程に起因するのがWISHE
  - 台風研究者A’
    - Aに似ているが、WISHEを海面過程が重要な定常状態にのみ適用し、Ooyama(1969)の数値計算で出てきた海面過程を含んだ発達をCooperative intensificationとみなす
  - 台風研究者B
    - 対流スケールと大規模な運動のスケールの相互作用に起因する場合、不安定の起源によらずCISKであり、WISHEはCISKに含まれる
  - 台風研究者C
    - 対流スケールと大規模な運動のスケールの相互作用がCISKの精神であることを認めつつ、相対的に大気を重視しているものをCISK, 海面過程を重視したものをWISHE

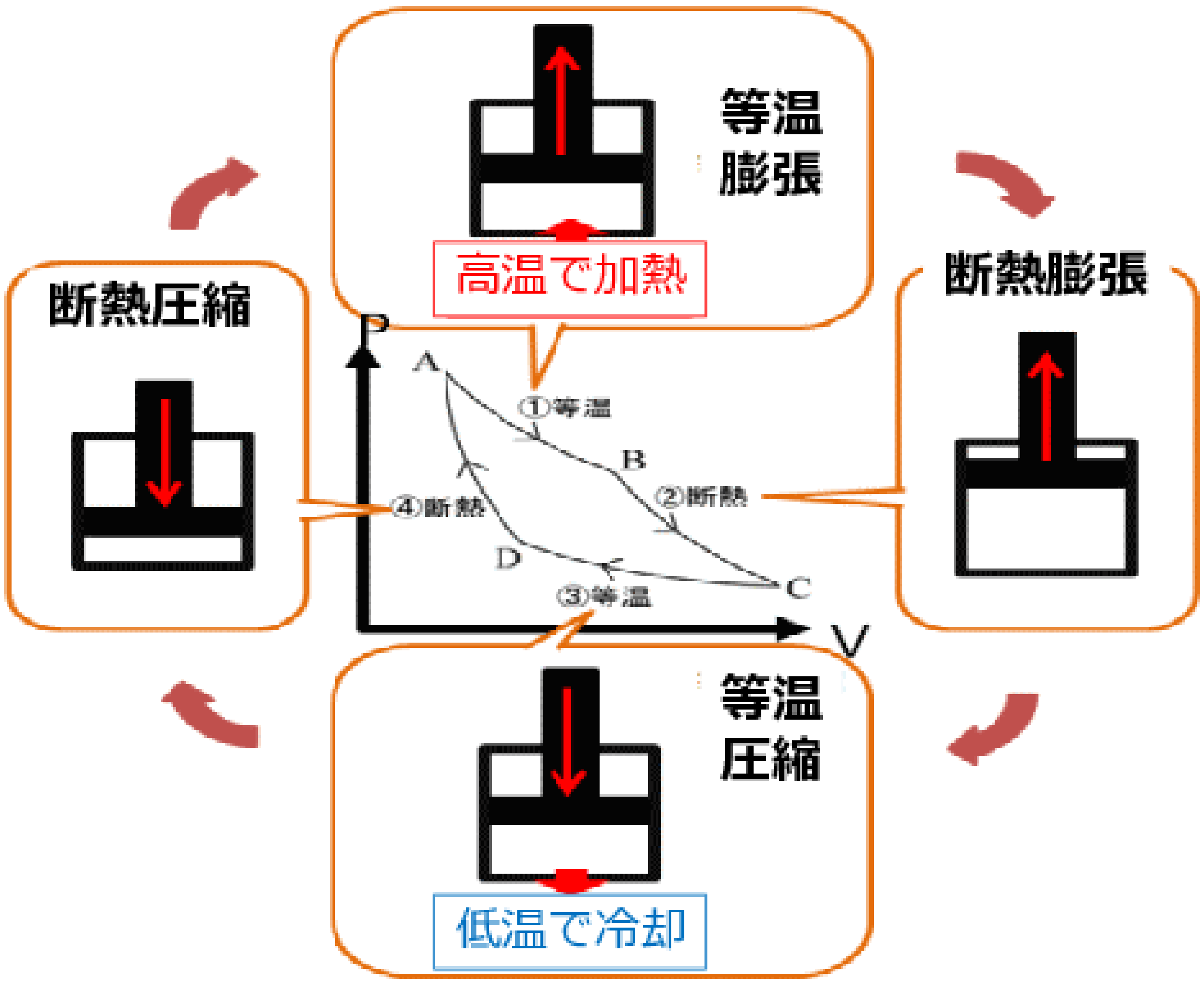
2種類のWISHE理論

- 簡便：定常状態における空気塊の状態変化が、熱力学的にカルノーサイクル「のような」経過をたどるとして最大風速を求める。

- 正式：定常状態において、自由大気中の壁雲領域で傾度風平衡・静水圧平衡・傾斜湿潤対流に対する中立性を仮定し、境界層と外出流域を境界条件とみなすことで、最大風速・中心気圧・台風の構造を求める。

カルノーサイクルエンジン

- 気体に(1)高温で加熱→(2)断熱膨張で温度を下げる→(3)低温で冷却→(4)断熱圧縮で温度を上げる、という操作を施すと最大の熱効率で加熱と冷却の差が仕事に転化される



湿潤比エントロピー：パーセルの持つエネルギー

- 湿潤比エントロピー $Tds_m = C_p dT - \alpha dp + Ldq_v$
- B→CとD→Aは湿潤比エントロピーを変えないという意味で「断熱過程」（B→Cで相変化が起こるがこれはLdq$_v$がC$_p$dT-αdpに変わること）

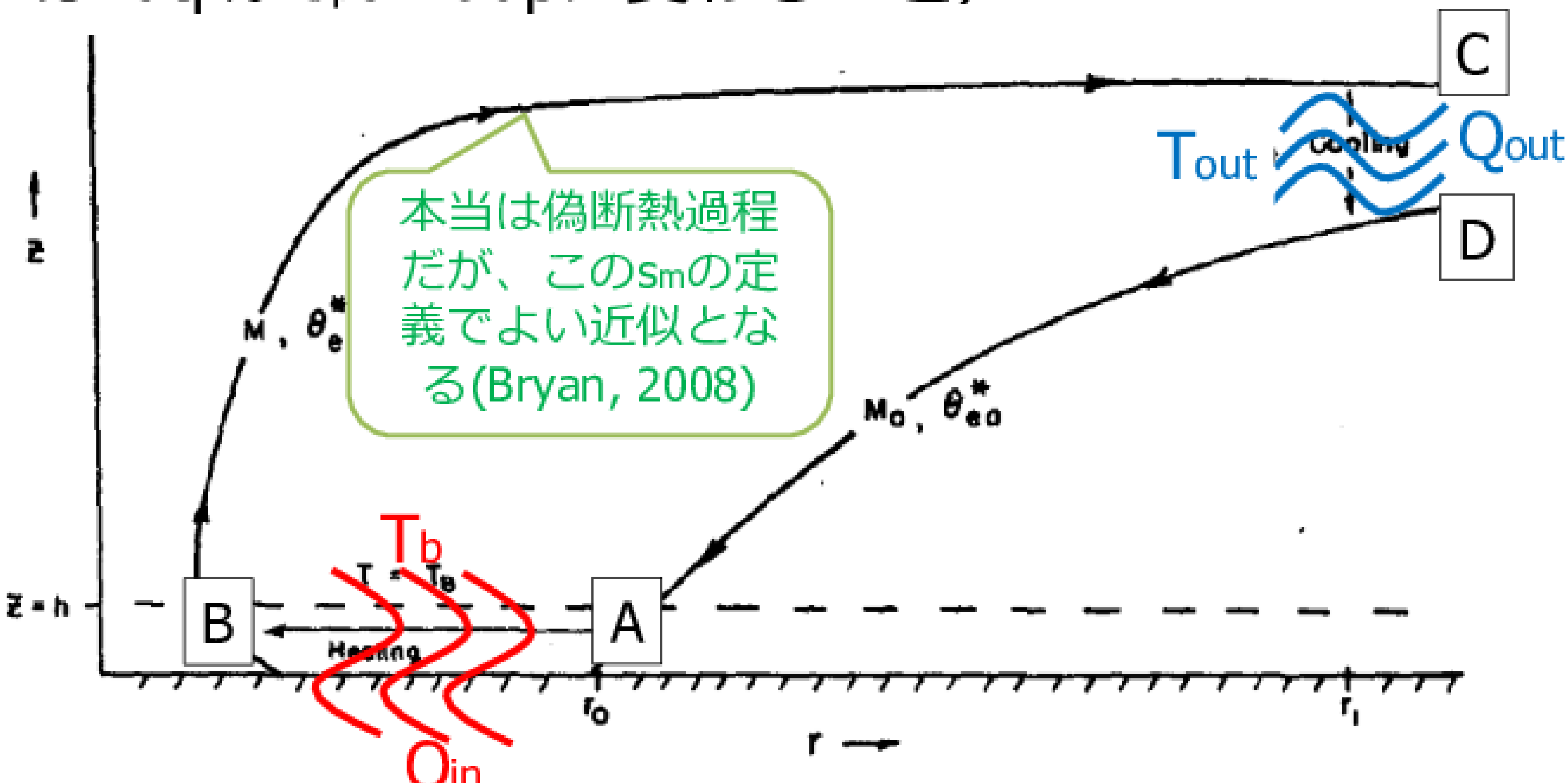


FIG. 13. The tropical cyclone as a Carnot heat engine. See text for explanation.

カルノーサイクルエンジンとしての台風

- カルノーサイクルエンジンでは、「加熱」と「仕事」の比が暖かい部分と冷たい部分の温度だけで熱効率が決まる。

$$\eta = \frac{W}{Q_{in}} = \frac{T_b - T_{out}}{T_b}$$

- 定常状態においては、仕事として得られる運動エネルギーが摩擦によって失われる分に等しいとして

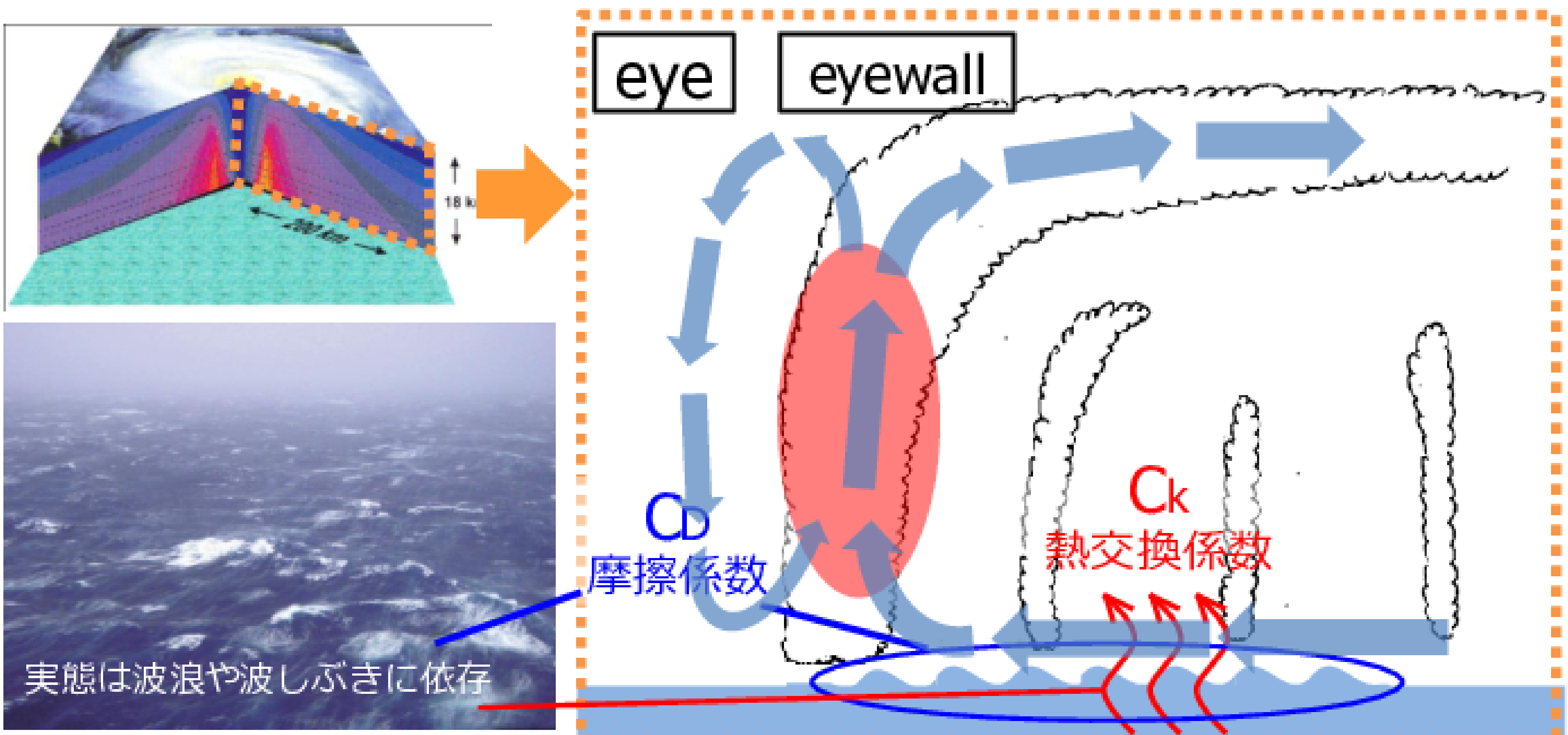
$$\frac{W}{Q_{in}} = \frac{\rho C_D v^3}{\rho C_k v \left(s^*_{m,sfc} - s_{m,b}\right) T_b}$$

- 上の2つの式を結び付けて

$$\boxed{v^2 = \frac{C_k}{C_D}\left(s^*_{m,sfc} - s_{m,b}\right)\left(T_b - T_{out}\right)}$$

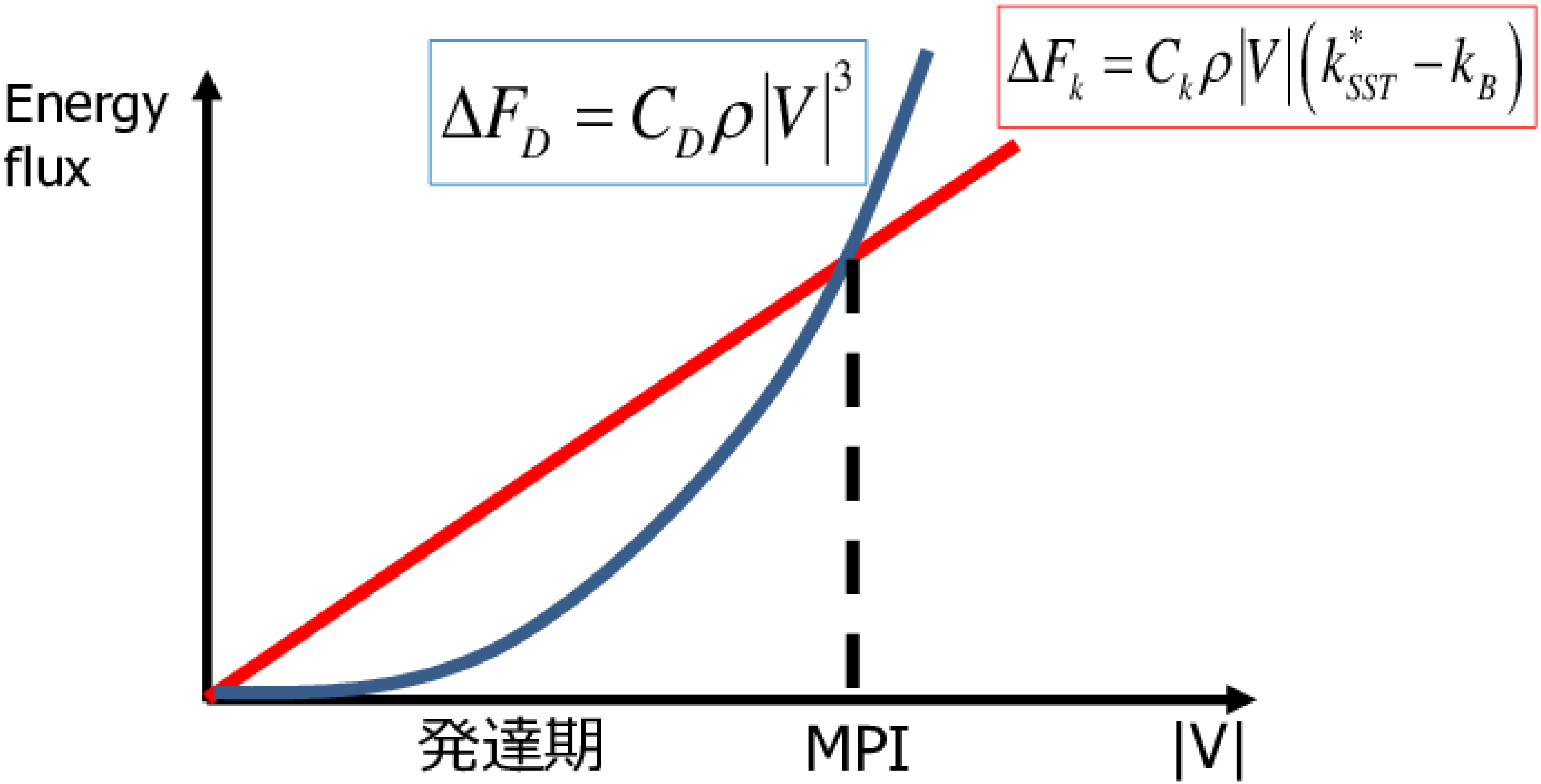
MPIを解釈する

$$v^2 = \frac{C_k}{C_D}\left(T_b - T_{out}\right)\left(s^*_{SST} - s_b\right)$$



発達期と成熟期の台風 (Wang, 2012)

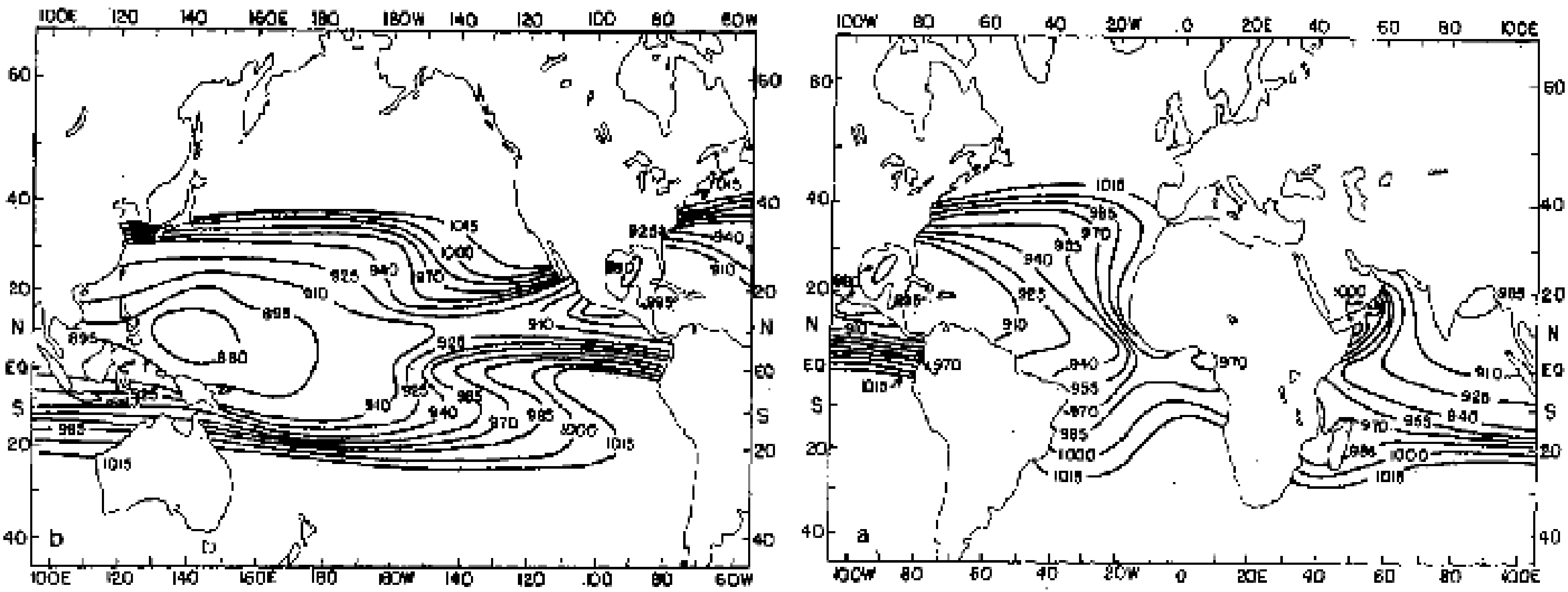
- 仕事に転化される熱エネルギーは風速に比例し、摩擦で大気が失うエネルギーは風速の3乗に比例。
- 風速が弱いとき摩擦も弱いので台風は発達できる。



講演後コメント：引用文献を訂正しました。

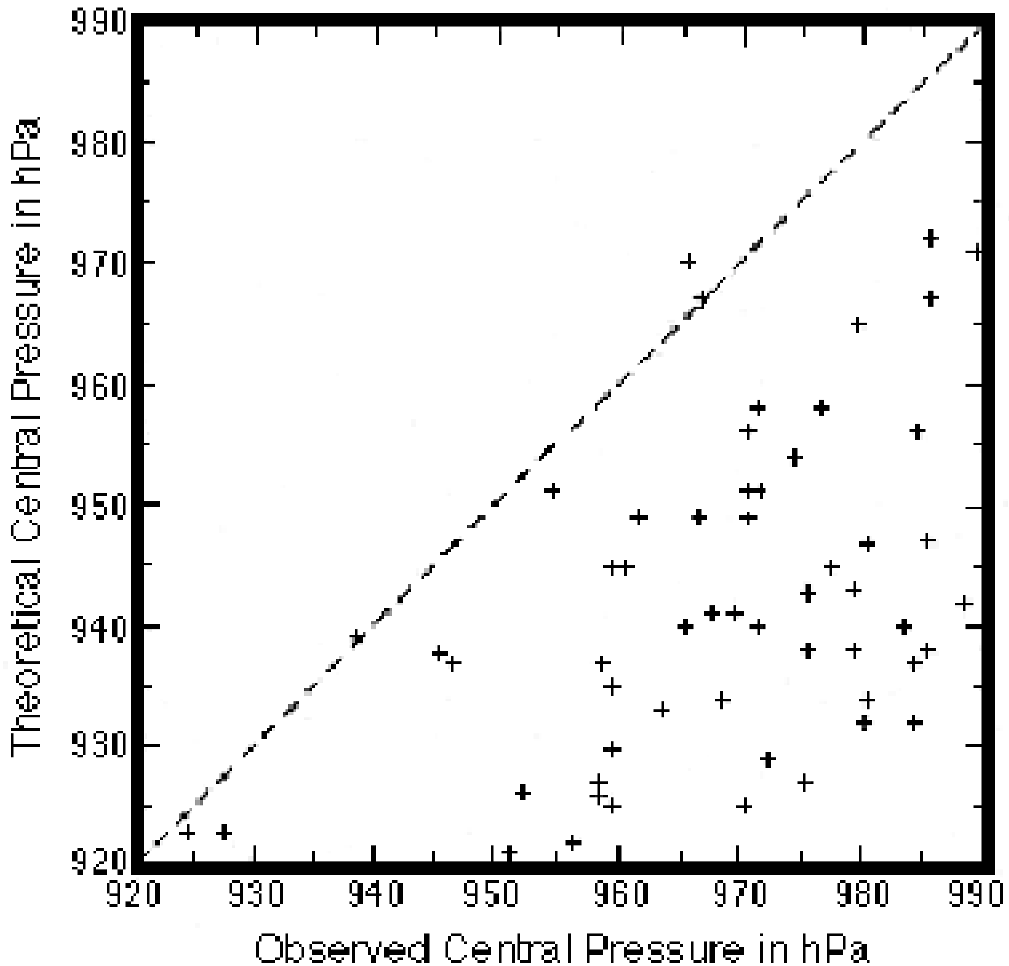
WISHE: 現実の台風強度との比較

- 9月の平均的な状況に対し、湿度78%として中心気圧に対するMPIを計算した。
- 北西太平洋や大西洋など、現実の台風が見られる地域でもMPIが低くなっている。



(Emanuel 1986)

MPI：現実大気の台風強度との比較



https://emanuel.mit.edu/hurricane-potential-intensity-maps-0

非静力学軸対称モデルとの比較

TABLE 2. Numerical experiments.

	Initial vortex					
Exp	v_{max} (m s^{-1})	r_m (km)	r_0 (km)	T_{surf} (°C)	Tropopause (mb)	Comments
A	12	82.5	412.5	26.3	100	Control run
B	2	82.5	412.5	26.3	100	Weak vortex
C	12	160.0	800.0	26.3	100	Large vortex
D	12	41.0	206.0	26.3	100	Small vortex $\Delta r = 7.5$ km, $r_{outer} = 750$ km, $l_H = 1500$ m
E	12	82.5	412.5	26.3	100	Dry to 30% RH above boundary layer
F	12	82.5	412.5	26.3	300	Increase tropopause
G	12	82.5	412.5	31.3	100	Increase SST
H	12	82.5	412.5	31.3	200	Increase SST and lower tropopause
I	12	82.5	412.5	31.3	300	Increase SST and lower tropopause
J	12	82.5	412.5	26.3	100	Limited Newtonian cooling $\|R\| < 2$ K d^{-1}
K	12	82.5	412.5	26.3	100	Zero Newtonian cooling $R = 0$

TABLE 3. Comparison of theory and numerical experiment.

	Numerical experiments*										Theory		
Exp	T_s (°C)	T_{out} (K)	$(T_{out})_{est}^{\dagger}$ (K)	T_B (K)	ϵ	r_{max} (km)	r_0 (km)	RH_{as} (%)	p_c (mb)	v_{max} (m s^{-1})	p_c (mb)	v_{max} (m s^{-1})	r_0 (km)
A	26.3	228	203	295.	.23	38	400	81.8	973	46	975	48	380
F	26.3	245	244	295.	.17	45	400	81.8	991	38	989	42	340
G	31.3	207	197	298.	.31	30	900	78.6	903	77	903	72	570
H	31.3	227	227	298.5	.24	40	750	78.6	937	66	944	63	490
I	31.3	246	246	298.5	.18	50	700	78.6	964	55	972	54	430
J	26.3	216	203	295.	.27	38	340	81.8	961	51	961	53	440
K	26.3	212	203	295.	.28	38	370	81.8	957	57	956	59	450

(Rotunno and Emanuel, 1987)

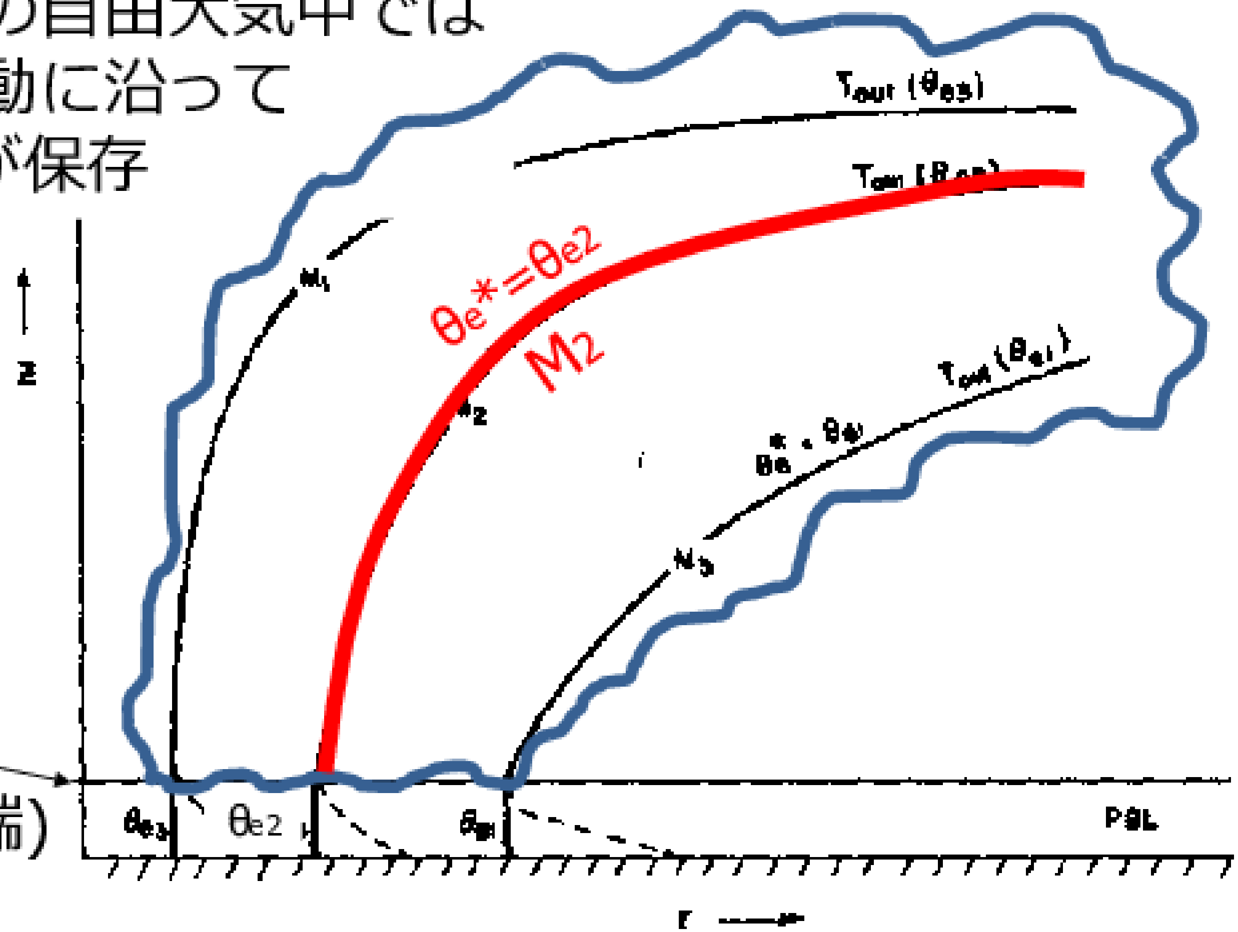
正式なWISHEの定式化

- 以下、Emanuel(1986)をかいつまんで説明。
- 自由大気
  - ➢傾度風平衡
  - ➢静水圧平衡
  - ➢傾斜湿潤対流に対する中立（＝等M面と等θ_e面が平行）
- 境界層
  - ➢鉛直1層のスラブモデルとして近似
  - ➢接線風は動径風に比べて十分に小さい
  - ➢湿度は一定。気温も一定。

  など

WISHEにおける自由大気のモデル化

内部コアの自由大気中では
気塊の運動に沿って
M, θe, smが保存



z=h
(境界層上端)

FIG. 1. Structure of the steady-state model. Curved lines above the planetary boundary layer (PBL) represent surfaces of constant angular momentum (M) and saturated equivalent potential temperature (θ_e^*). Solid lines in PBL show surfaces of constant θ_e while M is shown by dashed lines.

(Emanuel, 1986)

斜向(対称)不安定

- 鉛直方向の変位に関する安定性は成層安定度

$$N^2 = \frac{g}{\theta_0} \frac{\partial \theta}{\partial z}$$

- 水平方向の変位に関する安定性は慣性安定度

$$I^2 = \frac{1}{r^3} \frac{\partial M^2}{\partial r}$$

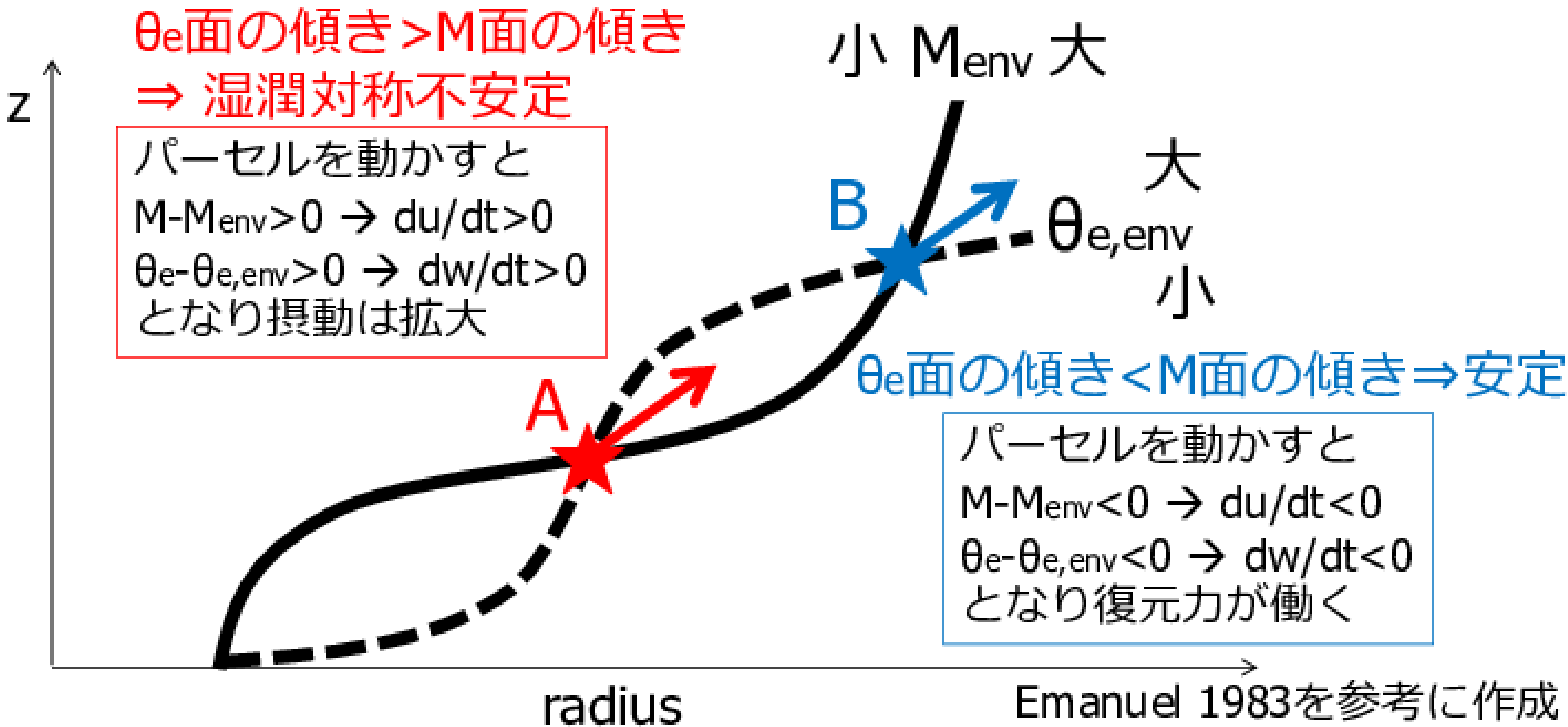
- 成層安定度と慣性安定度に関して、ともに安定であっても、斜向する変位に関して不安定ということがあり得る(斜向(対称)不安定)

湿潤斜向不安定（傾斜湿潤対流に対する安定性）

- $\theta_{e,env}$面の傾き＞M_{env}面の傾き⇒斜向に対して摂動が増幅 (Emanuel, 1983; Schecter and Montgomery, 2007)。

uとwの方程式(w>0)

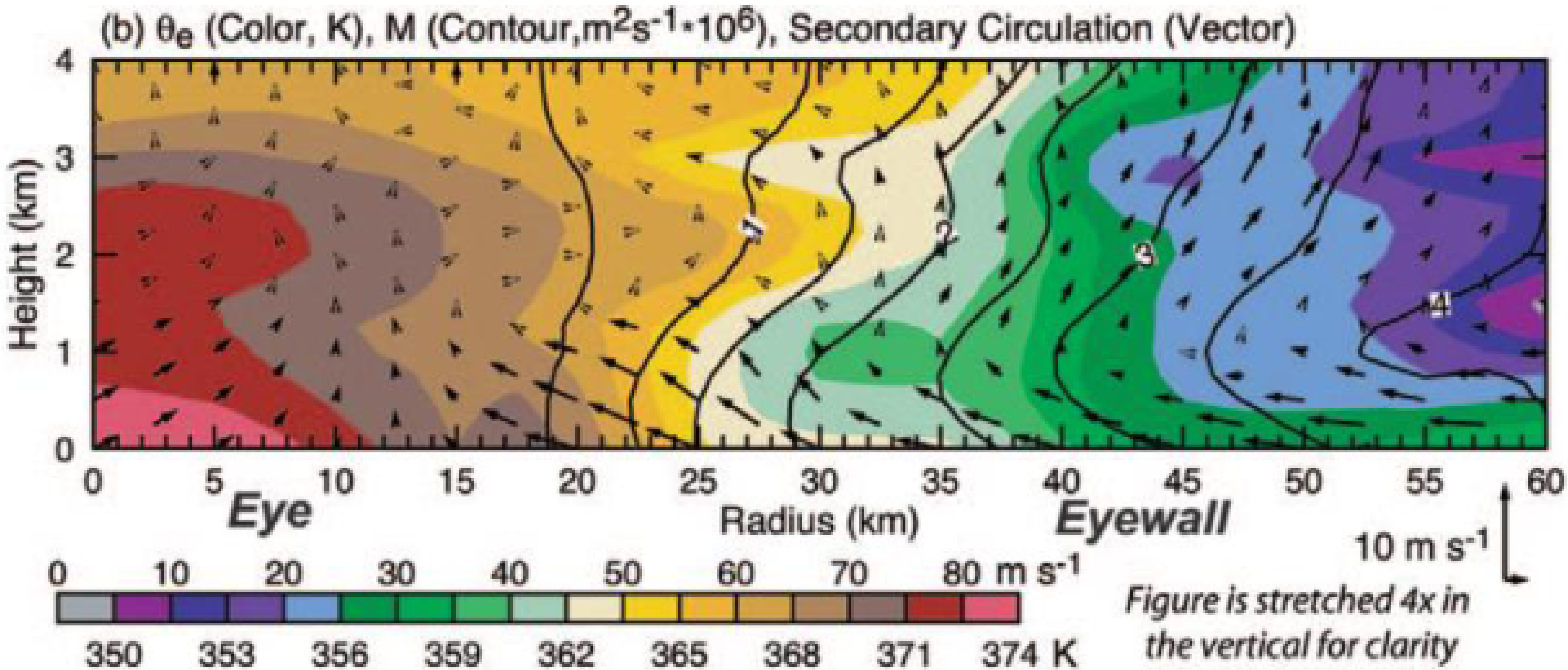
$$\frac{\partial u}{\partial t}=\left(f+\frac{2\bar{v}}{r}\right)\left(\frac{M-M_{env}}{r}\right),\quad \frac{\partial w}{\partial t}=\frac{g}{\theta_{e0}}\left(\theta_e-\theta_{e,env}\right)$$



Emanuel 1983を参考に作成

傾斜湿潤対流平衡: 等θe面と等M面が平行

- 等θe面(=等sm面)と等M面が平行であることは傾斜湿潤対流(湿潤斜向不安定の条件)に対して中立であることを意味している(Emanuel, 1983; Schecter and Montgomery, 2007)。
- 台風の発達の時間スケール（～1day）が傾斜湿潤対流の時間スケール（～1h）に比べて十分に長いため、台風の発達を考える上では、不安定が解消されて中立だと想定している。



(Montgomery et al., 2006)

自由大気のモデル化

仮定1

傾度風平衡 $\alpha\left(\frac{\partial p}{\partial r}\right)_z=\frac{v^2}{r}+fr$

仮定2

静水圧平衡 $\alpha\left(\frac{\partial p}{\partial z}\right)_r=-g$

絶対角運動量の定義 $M=\frac{1}{2}fr^2+rv$

座標変換

$$\alpha\left(\frac{\partial p}{\partial r}\right)_z=\frac{M^2}{r^3}-\frac{1}{4}f^2r$$

$$g\left(\frac{\partial z}{\partial p}\right)_r=-\alpha$$

偏微分の公式 $\left(\frac{\partial p}{\partial r}\right)_z=-\left(\frac{\partial p}{\partial z}\right)_r\left(\frac{\partial z}{\partial r}\right)_p$

rで微分

$$g\left(\frac{\partial z}{\partial r}\right)_p=\frac{M^2}{r^3}-\frac{1}{4}f^2r$$

pで微分

温度-風関係

$$\frac{1}{r^3}\left(\frac{\partial M^2}{\partial p}\right)_r=-\left(\frac{\partial \alpha}{\partial r}\right)_p$$

αをp,s_mの関数とみなす $\rightarrow$ $\left(\frac{\partial \alpha}{\partial r}\right)_p=\left(\frac{\partial \alpha}{\partial s_m}\right)_p\left(\frac{\partial s_m}{\partial r}\right)_p$ $\rightarrow$

温度-風関係 $\frac{1}{r^3}\left(\frac{\partial M^2}{\partial p}\right)_r=-\left(\frac{\partial \alpha}{\partial s_m}\right)_p\left(\frac{\partial s_m}{\partial r}\right)_p$

湿潤比エントロピーの定義

$$T\delta s_m = C_p \delta T + L\delta q_v - \alpha\delta p$$

湿潤比エンタルピーの定義

$$\delta k_m = C_p \delta T + L\delta q_v = T\delta s_m + \alpha\delta p$$

$$\left(\frac{\partial k_m}{\partial p}\right)_{s_m} = \alpha, \left(\frac{\partial k_m}{\partial s_m}\right)_p = T$$

熱力学におけるMaxwellの関係式

$$\left(\frac{\partial \alpha}{\partial s_m}\right)_p = \left(\frac{\partial T}{\partial p}\right)_{s_m}$$

温度-風関係

$$\boxed{\frac{1}{r^3}\left(\frac{\partial M^2}{\partial p}\right)_r = -\left(\frac{\partial \alpha}{\partial s_m}\right)_p \left(\frac{\partial s_m}{\partial r}\right)_p}$$

$$\frac{1}{r^3}\left(\frac{\partial M^2}{\partial p}\right)_r = -\left(\frac{\partial T}{\partial p}\right)_{s_m} \left(\frac{\partial s_m}{\partial r}\right)_p$$

偏微分の公式

$$\left(\frac{\partial M}{\partial p}\right)_r + \left(\frac{\partial M}{\partial r}\right)_p \left(\frac{\partial r}{\partial p}\right)_M = 0$$

$$\left(\frac{\partial r}{\partial p}\right)_r \left(\frac{\partial M}{\partial r}\right)_p = \frac{r^3}{2M}\left(\frac{\partial T}{\partial p}\right)_{s_m} \left(\frac{\partial s_m}{\partial r}\right)_p$$

仮定3

等M面と等sm面が平行 $\left(\frac{\partial s_m}{\partial r}\right)_p = \frac{ds_m}{dM}\left(\frac{\partial M}{\partial r}\right)_p$

$$\left(\frac{\partial r}{\partial p}\right)_M = \frac{r^3}{2M}\frac{ds_m}{dM}\left(\frac{\partial T}{\partial p}\right)_{s_m}$$

壁雲を通るある気塊のトラジェクトリについて、
境界層上端から外出流までをM(sm)に沿って、
pで積分して整理。rout>>rbであるとする。

壁雲を通る気塊に当てはまる関係式

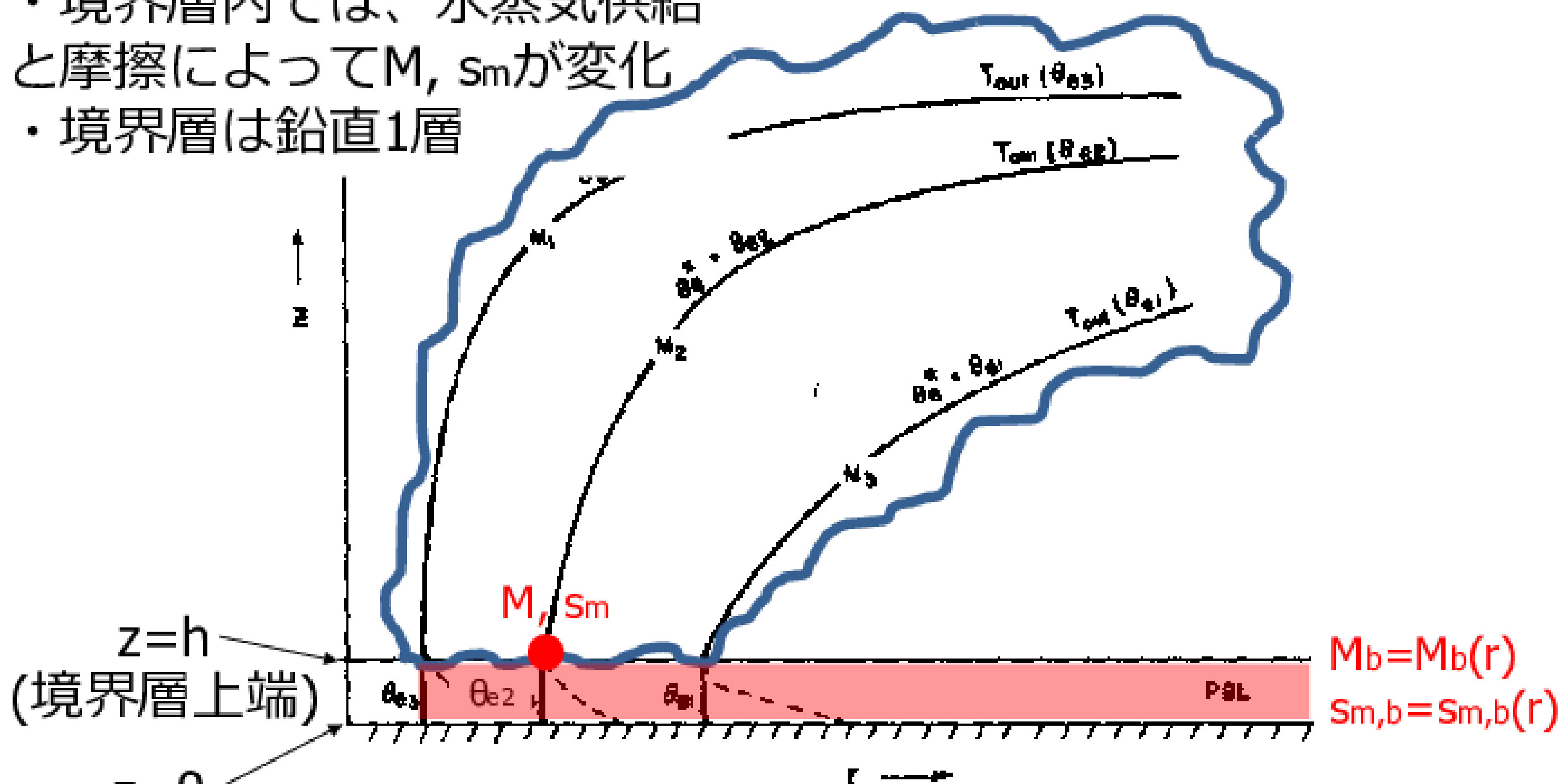
$$\boxed{v^2 = -M\frac{ds_m}{dM}\left(T_b - T_{out}\right)}$$

$v^2 = -M\dfrac{ds_m}{dM}(T_b - T_{out}) = -M\dfrac{(ds_m/dr)}{(dM/dr)}(T_b - T_{out})$の解釈

- 中心と外側の相当温位勾配ds_m/drがきついほど、接線風速は強くなる。
  （中心付近が暖かく湿っている）
- 中心と外側の絶対角運動量勾配dM/drがゆるいほど、接線風速は強くなる。
  （大きい絶対角運動量が中心付近まで輸送されている）
- 外出流の気温T_{out}が低いほど接線風速は強くなる。

境界層のモデル化(境界層上端におけるds_m/dM)

・境界層内では、水蒸気供給と摩擦によってM, s_mが変化
・境界層は鉛直1層



1. Structure of the steady-state model. Curved lines above the ry boundary layer (PBL) represent surfaces of constant angular momentum (M) and saturated equivalent potential temperature (θ_e^*). Solid lines in PBL show surfaces of constant θ_e while M is shown by dashed lines.

(Emanuel, 1986)

境界層のモデル化(境界層上端におけるds_m/dM)

鉛直1層としたときのMとs_mの方程式(上端で上昇流を仮定)

(Smith et al. 2008を一部改変)

$$\frac{u_b}{r}\frac{dM_b}{dr} = \underbrace{\cancel{\frac{w_c}{h}\left(v_b - v\right)}}_{\text{自由大気との乱流混合}} \underbrace{- \frac{C_D}{h}\left(\cancel{u_b^2} + v_b^2\right)^{1/2} v_b}_{\text{海面摩擦}}$$

$$u_b \frac{ds_{m,b}}{dr} = \underbrace{\cancel{\frac{w_c}{h}\left(s_{m,b} - s_{m,free}\right)}}_{\text{自由大気との乱流混合}} + \underbrace{\frac{C_k}{h}\left(\cancel{u_b^2} + v_b^2\right)^{1/2}\left(s^*_{m.sfc} - s_{m,b}\right)}_{\text{海面からの水蒸気・顕熱供給}} + \underbrace{\cancel{\delta}}_{\text{放射などの熱源}}$$

$$\downarrow$$

$$-\frac{ds_m}{dM} \approx -\frac{ds_{m,b}}{dM_b} = \frac{ds_{m,b}/dr}{dM_b/dr} = \frac{1}{M_b}\frac{C_k}{C_D}\left(s^*_{m.sfc} - s_{m,b}\right)$$

$$\downarrow$$

$$\boxed{v^2 = \frac{C_k}{C_D}\left(s^*_{m.sfc} - s_{m,b}\right)\left(T_b - T_{out}\right) = \frac{T_b - T_{out}}{T_b}\frac{C_k}{C_D}\left(k^*_{m.sfc} - k_{m,b}\right)}$$

おまけスライド

湿潤斜向不安定

$$\frac{\partial u}{\partial t} = \left(f + \frac{2\bar{v}}{r} \right) \left(\frac{M - M_{env}}{r} \right)$$

$$\frac{\partial}{\partial t} \left(\frac{M - M_{env}}{r} \right) = -\left(f + \bar{\zeta} \right) u$$

- なので結局、時間スケールは2π/Iになる。

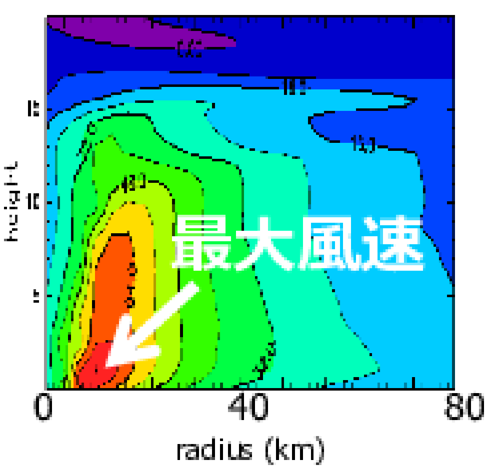
おまけ：超傾度風(傾度風平衡からの偏差)

- 軸対称モデルでは、最大接線風速は壁雲近傍の下層に現れ、傾度風平衡から期待される値よりも大きい (超傾度風) 。
- 壁雲近傍で吹き込みが止まったところに対応する。

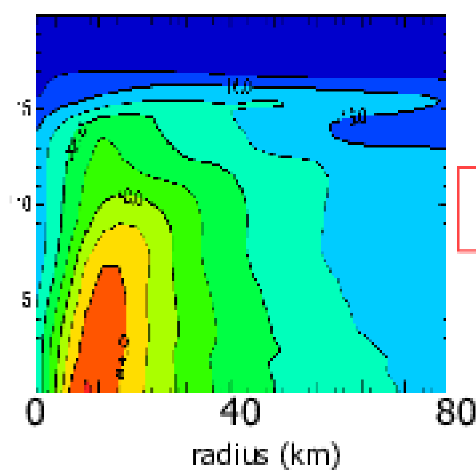
$$\underset{A}{\frac{Du}{Dt}} - \underset{B}{\frac{v^2}{r}} - \underset{C}{fv} = \underbrace{-\frac{1}{\rho}\frac{\partial p}{\partial r}}_{D} + \underset{E}{F_r}$$

- 観測にも見られる(光田1988, Montgomery et al. 2006)

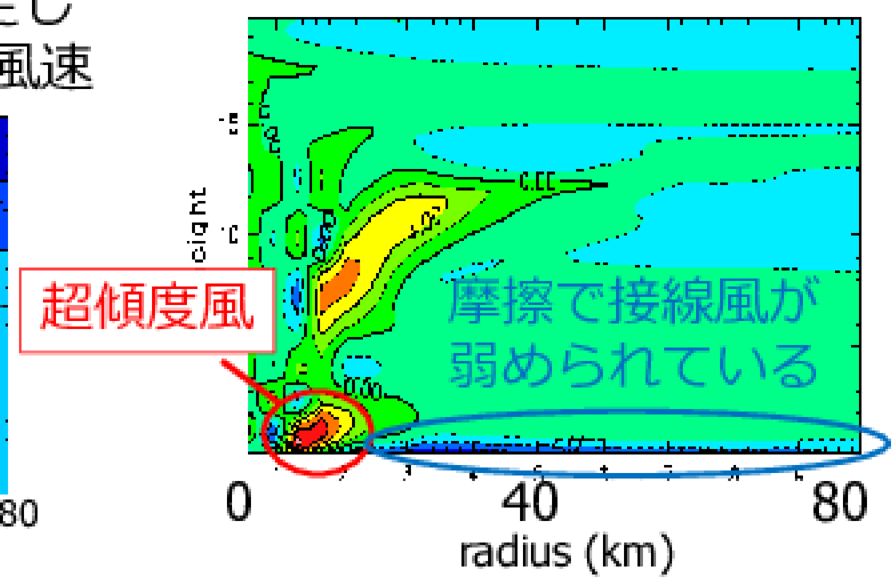
(a)接線風速



(b)傾度風平衡を仮定し
気圧から求めた接線風速



(c) (a)から(b)を引いたもの



Mitsuta (1988, JMSJ)

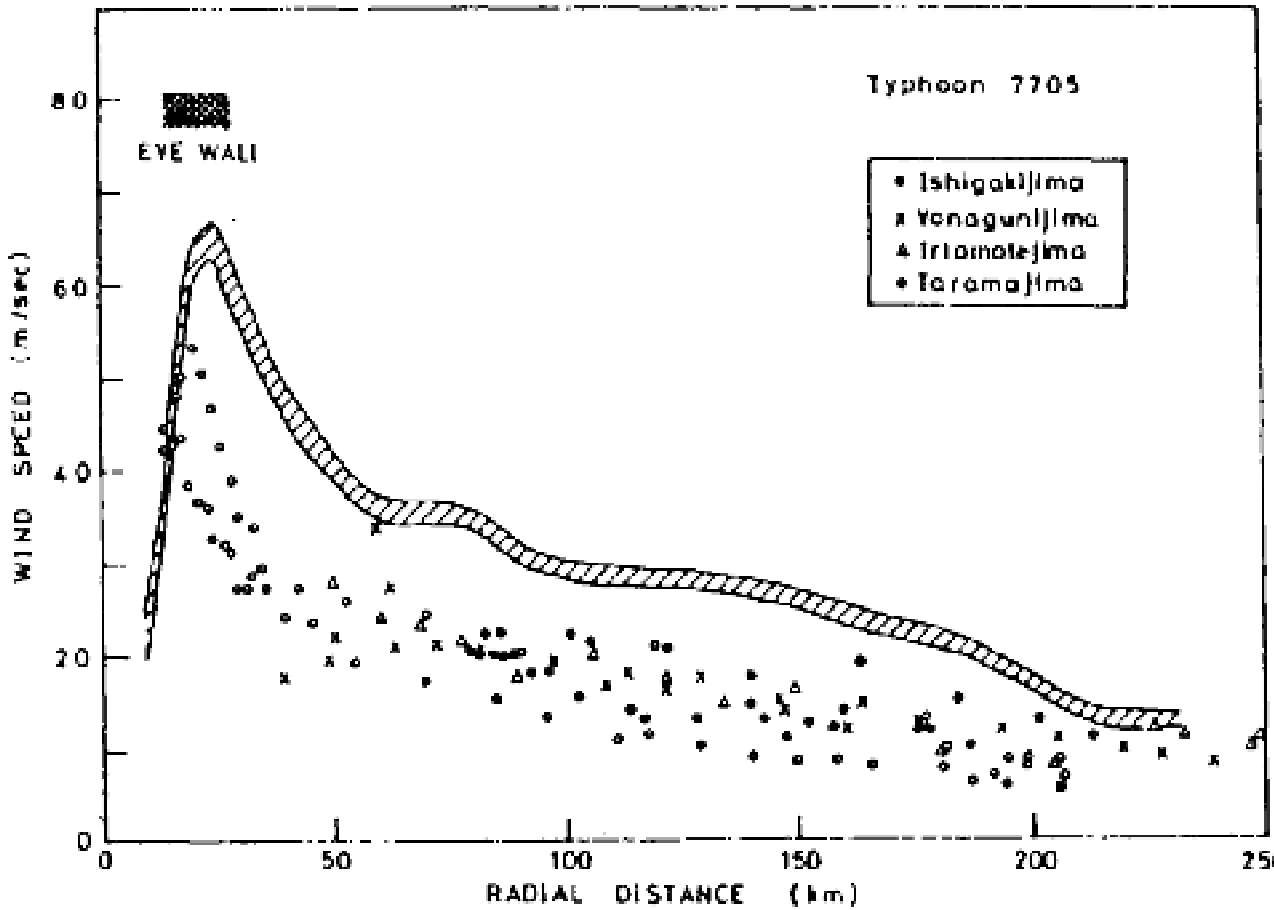
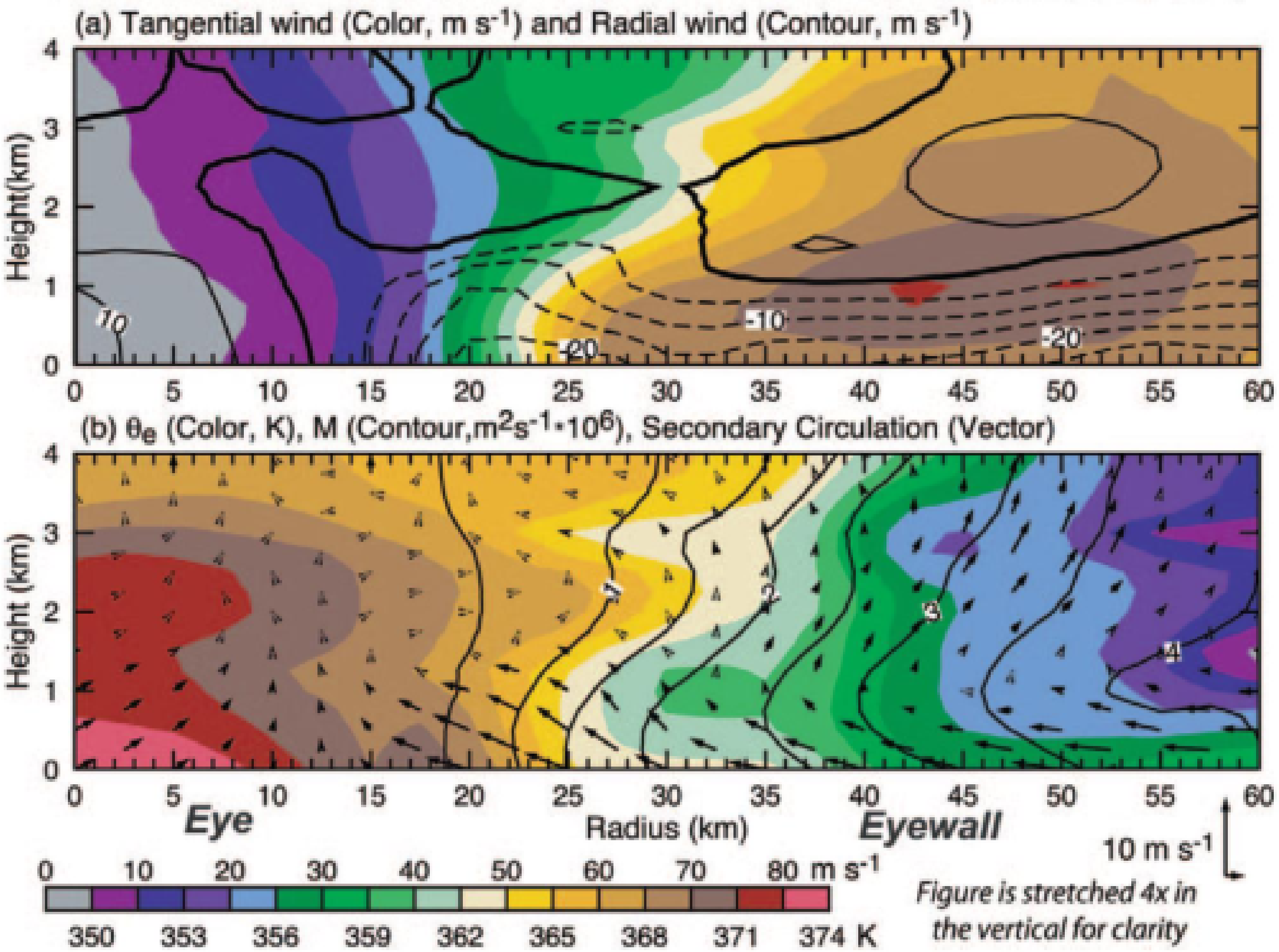


Fig. 3. Gradient and observed surface wind distribution as a function of relative distance from the typhoon center for Typhoon 7705, Vera. The gradient wind is shown by a curved line and surface winds by small marks. The shaded part of the gradient wind is the margin for typhoon translation as explained in the text.

超傾度風と最大風速

- 接線風速の最大値(r=42.5km, z=1km)はほぼ動径風速がゼロの地点に現れる。



湿潤PVを使った解釈(Emaneul, 1986)

- 絶対渦度ベクトルと絶対角運動量面は平行(証明済)。

$$\boldsymbol{\omega} \cdot \nabla M = 0$$

- 湿潤PVなる量を定義する

$$q_e = \frac{\boldsymbol{\omega} \cdot \nabla \theta_e}{\rho}$$

- 傾斜湿潤対流に対する中立性は絶対角運動量面と相当温位面が平行であることを意味するので、結局、

$$\boxed{q_e = 0}$$

- 壁雲の中では湿度100%、傾度風平衡・静水圧平衡を仮定すると、inversionにより、境界条件から流れ場が復元できる (=台風内部コアは境界条件で決まる) (?)

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