Fuelberg

- vertical motion ✓
- hydrostatic eq. ✓
- basic synoptic questions ✓
- severe WX outlook ✓
- convective outlook ✓
- Thermodynamic diagrams ✓
- various aspects of synoptic meteorology (I) ✓
- several basic principles of synoptic met. ✓
- " " " " (II) (III) ✓
- Convection related questions ✓
- gravity waves
- microburst/downburst
- sea/land breeze + slope/valley circulation
- land/sea breeze (I), (II)
- Atmospheric stability
- multicell + supercell storms (I)
- multicell + supercell storms (II) + Microbursts.
- dry lines.
- understanding of WX maps

(out of 40)

* vertical motion

✓ MET?: Fuelberg (45 min to 1 hour) **

- Diagnosis of large scale vertical motion is very important to both operational meteorologists and to researchers involved in synoptic studies.

Give as much information as you can about the following techniques for diagnosing vertical motion. Include material about the origins (derivation) and assumptions of each technique, a description of the procedures involved, strengths and weaknesses of each technique, and any other appropriate material to show your knowledge.

A. Kinematic method, including any commonly used adjustment techniques.
B. Quasi-geostrophic omega equation.
C. Adiabatic technique

A. Kinematic method (see Holton sec 3.5.2)

Recall cont. eq. $\frac{\partial u}{\partial x}+\frac{\partial v}{\partial y}+\frac{\partial w}{\partial z}=0$ (Cartesian coord.) or

$\left(\frac{\partial u}{\partial x}+\frac{\partial v}{\partial y}\right)_p+\frac{\partial \omega}{\partial p}=0$ (Isobaric coord.)

Integrate w.r.t. p $\rightarrow \omega(p)=\omega(p_0)-\int_{p_0}^{p}\left(\frac{\partial u}{\partial x}+\frac{\partial v}{\partial y}\right)_p dp$

consider a pressure-weighted vertical average of u, v (indicated by < >)

$\rightarrow \omega=\omega(p_0)+(p_0-p)\left(\frac{\partial\langle u\rangle}{\partial x}+\frac{\partial\langle v\rangle}{\partial y}\right)$

Note, can calculate $\frac{\partial u}{\partial x}, \frac{\partial v}{\partial y}$ from finite-diff. methods

Hence can determine vertical velocity from knowledge of horiz. winds. However, in normal mid-lat. winds. $\underline{V}\simeq \underline{V}_g$ where $\underline{V}_g$ is non-divergent. Hence, calculating DIV involves small diff. of nearly equal terms.

→ DIV is from small departures from geostrophic balance

Thus a 10% error in estimating one of the wind component → 100% error in DIV.

B. Quasi-geostrophic omega method (See Hoton sec 6.2.4)

Recall hydrostatic eq. $\frac{\partial p}{\partial z}=-\rho g=-\frac{g}{\alpha}=-\frac{p}{RT}g$

$\rightarrow -\alpha=-\frac{RT}{p}=\frac{g\partial z}{\partial p}=\frac{\partial\Phi}{\partial p}$ $\rightarrow T=-\frac{p}{R}\frac{\partial\Phi}{\partial p}$

Recall TDE

$\frac{\partial T}{\partial t}+u\frac{\partial T}{\partial x}+v\frac{\partial T}{\partial y}-S_p\omega=\frac{\dot{q}}{c_p}$ (Subst. for T)

$\rightarrow \frac{\partial}{\partial t}\left(-\frac{\partial\Phi}{\partial p}\right)+u\frac{\partial}{\partial x}\left(-\frac{\partial\Phi}{\partial p}\right)+v\frac{\partial}{\partial y}\left(-\frac{\partial\Phi}{\partial p}\right)-\overset{\rightarrow \sigma}{\boxed{S_p\frac{R}{p}}}\omega$

$=\frac{\dot{q}}{c_p}\left(\frac{R}{p}\right)=\frac{\dot{q}}{c_p}\left(\frac{\alpha}{T}\right)=\frac{\alpha}{c_p}\frac{\dot{q}}{T}=\frac{\alpha}{c_p}\frac{ds}{dt}$ ← entropy

Now we assume $\underline{V}=\underline{V}_g$ (geostrophic) + diabatic heating is small wrt other terms.

$\rightarrow \frac{\partial}{\partial t}\left(-\frac{\partial\Phi}{\partial p}\right)+\underline{V}_g\cdot\nabla_H\left(-\frac{\partial\Phi}{\partial p}\right)-\sigma\omega=0$: Approx. form of TDE

Also consider QG vorticity eq.

$\frac{\partial\zeta_g}{\partial t}=-\underline{V}_g\cdot\nabla(\zeta_g+f)+\overset{\rightarrow -f_0\nabla\cdot\underline{V}_H}{\boxed{f_0\frac{\partial\omega}{\partial p}}}$

Now we define geopotential tendency $\equiv\chi=\frac{\partial\Phi}{\partial t}$

Subst. into QG TDE + vorticity eq.

$\rightarrow$ TDE : $-\frac{\partial\chi}{\partial p}=-\underline{V}_g\cdot\nabla_H\left(-\frac{\partial\Phi}{\partial p}\right)+\sigma\omega$

vort. : $\frac{1}{f_0}\nabla^2\chi=-\underline{V}_g\cdot\nabla_H\left(\frac{1}{f_0}\nabla^2\Phi+f\right)+f_0\frac{\partial\omega}{\partial p}$

where $\zeta_g=\frac{1}{f_0}\nabla^2\Phi \longrightarrow \frac{\partial\zeta_g}{\partial t}=\frac{1}{f_0}\nabla^2\chi$

Now we eliminate χ between these eqs. → expression for ω

We take ∇^2 of TDE, $\frac{\partial}{\partial p}$ of vort, subtract eqs to eliminate χ term

$$\rightarrow \underbrace{\left(\nabla^2+\frac{f_0^2}{\sigma}\frac{\partial^2}{\partial p^2}\right)\omega}_{\text{Laplacian of }\omega}=\underbrace{\frac{f_0}{\sigma}\frac{\partial}{\partial p}\left[\underline{V}_g\cdot\nabla\left(\frac{1}{f_0}\nabla^2\Phi+f\right)\right]}_{\text{vert. shear of horiz. geostrophic vort. adv.}}+\underbrace{\frac{1}{\sigma}\nabla^2\left[\underline{V}_g\cdot\nabla\left(-\frac{\partial\Phi}{\partial p}\right)\right]}_{\text{Lap. of }\ominus\text{ horizontal temp. adv.}}$$

Assuming ω is a periodic function, $\nabla^2\omega\propto-\omega$

So, we have $-\omega\propto$ vert. shear of horiz. vort. adv. + hori. temp adv.

$\rightarrow \binom{\text{rising}}{\text{sinking}}$ motion $\propto$ rate of $\binom{\text{increase}}{\text{decrease}}$ with height of $\binom{+}{-}$ vort. adv.

$+\binom{\text{warm}}{\text{cold}}$ advection

Hence the vert. velocity field that results ensure temperature changes are hydrostatic, and vorticity changes are geostrophic. Note that the two terms, even though they may result from diff. physical processes, still tend to cancel to a large degree.

C. Adiabatic method (see Holton sec 3.5.3)

Recall first law of thermodynamic.

$c_p\frac{dT}{dt}-\alpha\frac{dp}{dt}=\dot{q}$, expand $\frac{dT}{dt}=\frac{\partial T}{\partial t}+u\frac{\partial T}{\partial x}+v\frac{\partial T}{\partial y}+\omega\frac{\partial T}{\partial p}$ $\omega \rightarrow \frac{dp}{dt}$

$\rightarrow c_p\left(\frac{\partial T}{\partial t}+u\frac{\partial T}{\partial x}+v\frac{\partial T}{\partial y}+\omega\frac{\partial T}{\partial p}\right)-\alpha\omega=\dot{q}$

$$\rightarrow \frac{\partial T}{\partial t}+\underbrace{u\frac{\partial T}{\partial x}+v\frac{\partial T}{\partial y}}_{\text{horiz. temp. adv.}}-\omega\underbrace{\left(-\frac{\partial T}{\partial p}+\frac{\alpha}{c_p}\right)}_{\text{static stability parameter}\equiv S_p}=\frac{\dot{q}}{c_p}$$ → Thermodynamic energy eq.

If we assume diabatic term ($\dot{q}$) is small wrt other terms, we can solve for ω.

$\longrightarrow \omega=\frac{1}{S_p}\left(\frac{\partial T}{\partial t}+u\frac{\partial T}{\partial x}+v\frac{\partial T}{\partial y}\right)$ ✓

we can measure static stability + horiz. adv. fairly well. Here the problem is the local change of temp. unless obs. are taken at close intervals in time. It is very difficult to estimate $\frac{\partial T}{\partial t}$ very accurate over a wide area.

⟶ See next page ?

- Vorticity method

Start with QG vorticity eq.

$$\frac{\partial \zeta_g}{\partial t} = -\mathbf{V}_g \cdot \nabla(\zeta_g + f) - f_0 \nabla \cdot \mathbf{V}$$

Recall cont. eq $\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial \omega}{\partial p} = \nabla_p \cdot \mathbf{V} + \frac{\partial \omega}{\partial p} = 0$

$\rightarrow \frac{\partial \zeta_g}{\partial t} = -\mathbf{V}_g \cdot \nabla(\zeta_g + f) + f_0 \frac{\partial \omega}{\partial p}$

$\rightarrow \frac{\partial \omega}{\partial p} = \frac{1}{f_0}\left(\frac{\partial \zeta_g}{\partial t} + \mathbf{V}_g \cdot \nabla(\zeta_g + f)\right)$

→ We can integrate using B.C. (eg $\omega = 0$ @ $p = 0$, $p = p_s$)

→ eq. of ω

Hence we must evaluate ① Local change of ζ_g and ② Horiz. vort. adv.

Note : $\frac{\partial \zeta}{\partial t}$ is difficult to estimate accurately as you need at least two time measurements at each point to estimate the time rate of change.

✓ Satellite I.R method (using Cloud top temperature changes)

$$\frac{DT_{BB}}{Dt} = \frac{\partial T_{BB}}{\partial t} + \mathbf{V} \cdot \nabla T_{BB} + W \frac{\partial T_{BB}}{\partial z}$$

Horizontal adv. + local changes neglected ($T_{BB} = T_{air} = T$)

$$W = \frac{\frac{DT_{BB}}{Dt}}{\frac{\partial T}{\partial z}}$$

where lapse rate $\frac{\partial T}{\partial z}$ is found from soundings and $\frac{DT_{BB}}{Dt}$ is found from time sequence of IR satellite photos

Assume : Changes in T_{BB} at cloud tops are due to vertical motion only. Air below moves upward w/ the cloud.

- Advantages / disadvantages

(A) kinematic method

Advantages : ① ease of use → only need vertical profile of u, v

② O'Brien correction at tropopause → B.C. Cont. eq. satisfied

Disadv. : ✓① No info. from mass field directly considered

② DIV is small diff. between large quantities → small errors in wind field → large errors in DIV, ω

(B) QG ω-eq

Adv. : ① corresponds to geostrophic, hydrostatic atmosphere

② only need measurement at single time (no $\frac{\partial}{\partial t}$ term) — diagnostic

③ good for qualitative understanding within QG framework

Disadv : ① Difficult to evaluate quantitatively (∇^2 terms, etc)

② Large degree of cancel between vort. adv + temp. adv. terms.

(C) Adiabatic method

Adv : ① Incorporates both mass + wind fields into calc.

② Simple calc. mathematically

Disadv : ① require $\frac{\partial T}{\partial t}$ → must have mult. meas. of T in time to calc. ω

② Diabatic heating difficult to measure (assume = 0. ignores effects sometimes impossible)

(D) Vorticity method

Adv : ① Use of calc. (only requires wind field info → ζ)

✓② Largescale motion fields in mid-lats are largely rotational (and non-div.)

③ No assumption requires for adiabatic vs diabatic

Disadv : ① require $\frac{\partial \zeta}{\partial t}$ → difficult to measure accurately (requires mult. time measurement)

② $\frac{\partial \zeta}{\partial t}$, $\mathbf{V} \cdot \nabla(\zeta + f)$ cancel to large degree → errors in ω can be large!

✓ Q-vectors : Rep. of the QG eq (combines terms)

Approx. ω eq → $\nabla^2 \omega \simeq -\frac{\partial}{\partial p}[\text{vort. adv.}] - \nabla^2[\text{temp. adv.}]$

$$-\omega \approx +\frac{\partial}{\partial z}[\text{vort. adv.}] + [\text{temp adv}]$$

$$= \underbrace{-2\nabla \cdot \mathbf{Q}}_{\text{combo of forcing}} - \underbrace{\frac{R}{\sigma p}\beta \frac{\partial T}{\partial x}}_{\text{adv. of earth's vort}}$$

$$\mathbf{Q} = -\frac{R}{\sigma p}\begin{bmatrix} \frac{\partial \mathbf{V}_g}{\partial x} \cdot \nabla_p T \\ \frac{\partial \mathbf{V}_g}{\partial y} \cdot \nabla_p T \end{bmatrix} = \left[\begin{matrix} Q \\ Q \end{matrix}\right.$$

Neglecting 2nd term → CONV(DIV) of Q → rising (sinking) motion!

How to eval. on WX map? Define local coord. where x($\hat{i}$) lies along isotherm (cold air left)

$\rightarrow \frac{\partial T}{\partial x} \equiv 0 \longrightarrow \nabla_p T = \frac{\partial T}{\partial y}$ (always < 0 as cold air to left)

$|\mathbf{Q}| = -\frac{R}{\sigma p}\left|\frac{\partial T}{\partial y}\right|\left(\mathbf{k} \times \frac{\partial \mathbf{V}_g}{\partial y}\right) = +\frac{R}{\sigma p}\left|\frac{\partial T}{\partial y}\right|\frac{\partial \mathbf{V}_g}{\partial x}$, div $-\mathbf{k} \times \frac{\partial \mathbf{V}_g}{\partial x}$ → 90° to right of $\frac{\partial \mathbf{V}_g}{\partial x}$.

Eval. Q on either side (± increment in x direction) → determine CONV/DIV of Q → ω direction.

✓ - Kinematic method

Integrate eq. of cont. from top to bottom atmos.

$$\int_{\omega(p'=0)}^{\omega(p'=p)} \frac{\partial \omega}{\partial p'} dp' = -\int_0^p \delta \, dp'$$

$\delta = \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y}$: Horiz. div. $\delta + \frac{\partial \omega}{\partial p} = 0$, $\frac{\partial \omega}{\partial p} = -$

upper B.C. $\Rightarrow \omega(p'=0) = 0$ $\boxed{\omega(p) = -\int_0^p \delta \, dp'}$ ← kinematic vert. vel.

10% error in horizontal wind component (O(10m/s)) leads to ~1m/s error (instrument or sampling error). Errors in divergence(δ) could average ~20%. But a 1m/s error in wind ⇒ 100% error point estimate of ω would be inaccurate

"O'Brien's adjustment" ? → look at Stuart's question

We would expect the vertical velocity at the tropopause to be zero due to the high static stability and a sign reversal of the horizontal temperature gradient. If we let the upper B.C. $\Rightarrow \omega_{trop}$ + lower B.C $\Rightarrow \omega_0$, we can "adjust" the kinematic computation by an amount $\omega_{trop} - \hat{\omega}_{trop}$. We can then adjust $\hat{\omega}$ at all levels below the trop. assuming that the adjustment is zero at the ground. and increases linearly with height. This is equivalent to adjusting the estimates of divergence (δ) by an amount $\overline{\Delta\delta}$ which is independent of pressure

$$\omega(p) = \omega_0 - \int_{p_0}^{p} \hat{\delta} \, dp' - \int_{p_0}^{p} \overline{\Delta\delta} \, dp' = \hat{\omega} + K_{\delta_0}(p_0 - p)$$

where $\overline{\Delta\delta} \equiv K_{\delta_0} = \frac{\omega_{trop} - \hat{\omega}_{trop}}{p_0 - p_{trop}}$ $\boxed{K_{\delta_1} = 2\left(\frac{\omega_{trop} - \hat{\omega}_{trop}}{(p_0 - p_{trop})^2}\right)}$

For quadratic adjustment $\omega(p) = \omega_0 - \int_{p_0}^{p} \hat{\delta} \, dp' - \int_{p_0}^{p} 2\frac{(\omega_{trop} - \hat{\omega}_{trop})(p_0 - p)}{(p_0 - p_{trop})^2} dp'$

$$= \hat{\omega} + \frac{(\omega_{trop} - \hat{\omega}_{trop})(p_0 - p)^2}{(p_0 - p_{trop})^2}$$

The adjustments to $\hat{\omega}$ + $\hat{\delta}$ are minimized in a least-squares sense

* basic synoptic questions.

MET?: Fuelberg(30 min) * : multiple parts

- Question

a. You are analyzing height contours at 500 mb and have a geostrophic wind scale. The wind speed at a point having strongly cyclonic flow is 40 m/s. Based on that speed, your geostrophic scale suggests a contour spacing of ½ inch. Based on gradient wind considerations, should the spacing be more or less than ½ inch? Discuss your answer fully.

b. Draw streamlines and isotachs that demonstrate confluence and divergence existing simultaneously.

c. Draw a cross section that intersects a warm type occluded front. Label the regions “cold”, “coldest”, and “warm”.

d. Sketch a temperature-dew point sounding that represents a standard frontal inversion. Repeat the process showing a standard subsidence inversion.

e. Draw a diagram and explain the type of temperature advection that occurs when winds veer with altitude. Your discussion should be based on thermal wind concepts. → WAA

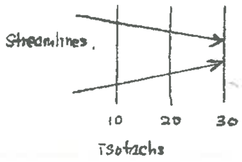
a) Gradient wind $\frac{V^2}{R} + fV = fV_g \longrightarrow \frac{V_g}{V} = 1 + \frac{V}{fR}$

For cyclonic curv., $R>0$, V, V_g defined to be always >0. so $\frac{V_g}{V} > 1$

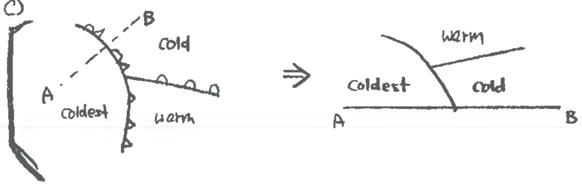
$\Rightarrow V_g > V$. As V_g is greater than observed wind, the spacing should be less than 1/2 inch.

gradient wind

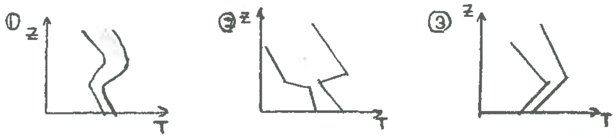
b)



c)



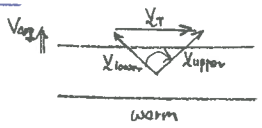
d) ① frontal inversion ② subsidence inversion + ③ nocturnal radiational inversion



e) Thermal wind $\rightarrow \frac{\partial \mathbf{V}_g}{\partial \ln p} = -\frac{R}{f}\mathbf{k}\times\nabla_p T$

$\mathbf{V}$ backing w/ height → cold adv.

$\mathbf{V}$ veering w/ 〃 → warm adv.



derive?

In p-coord,

$$\begin{cases} v_g = \frac{1}{f}\frac{\partial\Phi}{\partial x} \\ u_g = -\frac{1}{f}\frac{\partial\Phi}{\partial y} \end{cases}$$

$$\because \left(\frac{\partial\Phi}{\partial p} = -\alpha = -\frac{RT}{p}\right)$$

✓ $$\frac{\partial v_g}{\partial p} = \frac{1}{f}\frac{\partial}{\partial x}\left(\frac{\partial\Phi}{\partial p}\right) = -\frac{R}{f}\frac{\partial}{\partial x}\left(\frac{T}{p}\right) \quad \because p \to \text{const}$$

$$= -\frac{R}{fp}\frac{\partial T}{\partial x} \text{ —— ①}$$

✓ $$\frac{\partial u_g}{\partial p} = -\frac{1}{f}\frac{\partial}{\partial y}\left(\frac{\partial\Phi}{\partial p}\right) = \frac{R}{f}\frac{\partial}{\partial y}\left(\frac{T}{p}\right)$$

$$= \frac{R}{fp}\frac{\partial T}{\partial y} \text{ —— ②}$$

$$① \rightarrow \frac{\partial v_g}{\partial \ln p} = -\frac{R}{f}\left(\frac{\partial T}{\partial x}\right)_p$$

$$② \rightarrow \frac{\partial u_g}{\partial \ln p} = \frac{R}{f}\left(\frac{\partial T}{\partial y}\right)_p$$

⇓

vector form

$$\frac{\partial \mathbf{V}_g}{\partial \ln p} = -\frac{R}{f}\mathbf{k}\times(\nabla_H T)_p$$

Thermal wind : the vector difference between the geostrophic winds at 2 different levels.

* hydrostatic approx.

MET? : Fuelberg (?)

- Scale analysis of the third equation of motion for synoptic scale systems results in what is known as the hydrostatic approximation.

(a) Use the hydrostatic approximation to derive an expression for the height of the base of any atmospheric layer. ✓ ⟶ hypsometric eq.

(b) Use the expression derived in (a) to determine what effect warming in the lower stratosphere would have on existing cold-core 500 mb low.

(c) Use the expression derived in (a) to determine what effect cooling in the lower stratosphere would have on existing cold-core 500 mb low.

See Kloesel's question ⟶ same.

*Severe WX outlook

MET5511 (Mesomet): Fuelberg(1 hour) *

- You have gone up to the map room to see if any severe thunderstorms are likely to occur over the United States today, but, alas, the convective outlook chart prepared by the Severe Storms Center is not there. Some students are bewildered and wringing their hands because they don't have a sufficient background in synoptic meteorology to make their own outlook, but this is no problem for you because you taken plenty of these courses.
  a. Describe how you would prepare your own convective outlook chart. First discuss the concepts you would use to identify the region where general thunderstorms are likely. Then, discuss any additional steps that you would use to make the distinction between general and severe storms.
  b. Present this discussion in an organized manner. Be sure to describe the physical processes that you are seeking to identify. Do not merely describe the charts you will use.
  c. Remember that you are looking for general areas of potential storms. You are not trying to prepare a nowcast for a specific location.

- Parameters of interest in forming severe WX outlook.

* 1) 500mb PVA : Strong PVA lends upper air support to deep levels of rising motion for intense convection.
→ Coupled with low-level features.

2) Instability params : Various thermodynamic indices can indicate potential for intense convection (e.g. LI, TT, etc)

* 3) Mid (500mb) + upper-level (200-300mb) jets : Diffluent, divergent flow at this level supports low-level convergence + convection.

4) Low-level (850mb) jet : important LL (← low level) moisture convergence

※ For severe wx particularly looking for LF, RR of jetstreams.

5) LL dewpoint (850, Sfc) : LL source of moisture to feed convection

6) LL temp (850) : Look for warm adv.

* 7) 700mb dry intrusion : Lack of moisture at this level helps enhance convection (increases potential instability)
※ $-\frac{\partial}{\partial p}\overline{q'\omega'} > 0$ at top of BL.

8) Pressure/height changes : Sfc pressure falls, 500mb height falls
→ indicate dyn/thermodyn forcing conducive to severe WX
(ex : look for short wave moving into region → there can forecast organize convection)

9) Sfc analysis : Various features can help locate + focus convection (e.g. fronts, windshift lines, temp/moist. gradients)

✓ 10) height of wet-bulb zero : potential for damaging hail

→ see the next question

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* Convective outlook

MET5511C (Mesomet): Fuelberg (1 hour) **

- Thunderstorms occur under a variety of conditions. Assume that you walk into a forecast center during the morning and that the convective outlook for the afternoon hours is unavailable. Your task is to prepare such an outlook for the afternoon and early evening hours, i.e., a U.S. map showing areas of risk for either thunderstorms or severe thunderstorms.

Describe in detail the various procedures you would use to prepare your outlook. Include your use of the various 1200 GMT analyses and the numerical progs for 0000 GMT later in the day (the progs are available).

Besides a description of what you would do, explain WHY you would do it based on theoretical considerations. The WHY aspect is a very important part of this question.

This is your opportunity to give as much information as you can.

* How form a severe WX outlook?

Follow TR200 (Miller) checklist.

we need the following maps : sfc analysis, 850 heights, winds, temp, moisture
radar summary, 700 " " " "
500 " " " "
300/200 " " " "

Parameters of interest

① 500 mb PVA → by QG dynamics PVA at 500 mb is associated with low level convergence and rising motion. Strong upper level PVA lends upper air support to development of deep convection.

② Stability → various thermodynamic indices to access column stability. indicates potential for convection and suggest intensity of convection

③ Mid (500) + upper-level (300/200) jets → diffluent, divergent flow supports/triggers/enhances low level convergence + convection.

④ Low-level (850) jet → important for low-level moisture advection

⑤ Low-level (850 + sfc) dewpoint → low level moisture source to feed storms

⑥ 850 mb max temp. → low level warm advection

} focused region of warm, moist air.

⑦ 700 mb moist. change line → 700 mb temp advection
↳ 700 mb dry air intrusion can lead to more intense storm.

⑧ 12hr sfc pressure falls
500 mb height falls
→ pressure falls indicate dynamical/thermodynamic forcing which may be conducive to severe WX
↳ look for short wave advecting into region which can focus + organize convergence.

⑨ sfc analysis → location of fronts, temp/moisture gradients ; wind shift lines → All help locate and focus convection.

⑩ Height of wet bulb zero → potential for damaging hail.

← see the previous question.

★ Thermodynamic diagrams.

MET5511(Mesomet): Fuelberg(30 minutes)

- Thermodynamic diagrams are used very widely in synoptic meteorology. Describe the meanings of the following terms. I am not interested in how each is determined on the diagram, but, instead, what is the physical process underlying each term.

a. lifting condensation level
b. level of free convection
c. equilibrium level
d. positive areas
e. negative areas
f. convective condensation level
g. convective temperature
h. lid strength index
i. lifted index

a) LCL → lift parcel dry adiabatically to saturation. level where saturation occurs is LCL.

b) LFC → level at which a lifted parcel becomes positively buoyant and rises without external forcing.

c) EL → level at which parcel and environment are buoyantly neutral wrt each other

d) positive area → area between environmental sounding and that of a lifted parcel where the parcel temp exceeds the ambient temp.

e) negative area → positive area except ambient temp exceeds parcel temp.

f) CCL → extend saturation mixing ratio through dewpoint upward until intersect temp. profile. This is CCL.

g) convective temp → temperature a parcel must obtain so that it rises freely to the CCL

h) lid strength index $= \boxed{\underbrace{(\overline{\theta}_{sw}-\overline{\theta}_{w})}_{A}+\underbrace{(\overline{\theta}_{swL}-\overline{\theta}_{w})}_{B}}$

A ≡ buoyancy of a parcel lifted to where $\theta_{sw} = \overline{\theta}_{sw}$ ✓

B ≡ stabilizing effect of the lid ✓

$\overline{\theta}_{sw}$ ≡ average saturation wet bulb θ between the base of the lid and 500 mb.

$\overline{\theta}_{w}$ ≡ maximum wet bulb θ in the lowest 100 mb

$\overline{\theta}_{swL}$ ≡ maximum saturation wet bulb θ at base of lid

i) LI → identify a low level mean parcel. lift this parcel dry adiabatically to the LCL and then moist adiabatically to 500mb. LI = $T_{ambient}(500mb) - T_{parcel}(500$ Negative LIs indicate greater potential for convection provided a lifting mechanism. How you define the low level parcel obviously affects the LI. Commonly we take the mean T & Td in the lowest 50 mb of the sounding.

a) LCL

Lifting parcel dry-adiabatically to saturation. LCL is level sat. occurs.

b) LFC

Level at which lifted parcel becomes positively buoyant and rises without external forcing (to the EL).

c) EL

Level at which parcel is neutral buyant wrt environment.

d) Positive areas

Area between lifted parcel + env. sounding, where $T_{parcel} > T_{env}$.

e) negative areas

Same as positive areas, except $T_{parcel} < T_{env}$

f) CCL

Extend saturation mixing ratio from dewpoint upward until it intersects temp. profile. Level is CCL, which indicates level where condensation will occur due to upward motion.

g) Convective temp.

Temp. a parcel must obtain to rise freely to the CCL (follow CCL back down dry-adiabatically to the surface)

✓ h) Lid strength index

Index involving θ_w, θ_{sw}. Indicates buoyancy of parcel along with stabilizing effect of convective lid.

i) LI

Lift low-level mean parcel (eg. average of lowest 50 mb) dry-adiabatically to LCL, then moist adiabatically to 500 mb.

$LI = T_{env}(500mb) - T_{parcel}(500mb)$. Negative values of LI indicate buoyancy and hence greater potential for convection (assuming a lifting mech.)

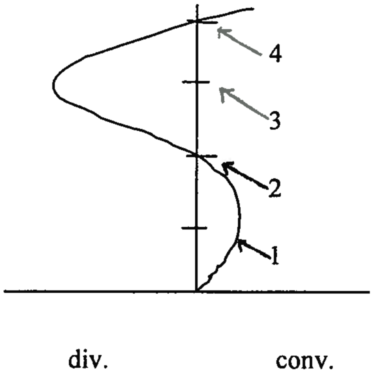
1/:

* Synoptic meteorology

MET?: Fuelberg(?)

- This question consists of several parts. It will test your knowledge about various aspects of synoptic meteorology.

A. Write the equation of continuity for an incompressible fluid (with *p* as the vertical coordinate). Explain in words what this equation tells us about the atmosphere. Integrate the equation to obtain an expression for kinematic vertical motion. Show all of your steps. The diagram below is a vertical profile of divergence. Using the equation you just derived (or your knowledge), sketch a corresponding profile of vertical motion. Label the axes "up" and "down" motion, and indicate on your diagram the altitudes that I have numbered 1, 2, 3 etc.

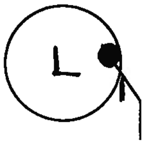


✓B. A cyclone at the surface has comparatively warm temperatures near its center. Will the cyclone become stronger or ✓weaker as altitude increases? Explain your answer thoroughly using the concept of thickness.

C. If winds change direction in a counterclockwise manner with increasing height, what type of temperature advection is occurring? Fully explain your answer with diagrams and a discussion. CAA

✓D. What is equivalent potential temperature? Why is equivalent potential temperature used often in meteorology? A layer of air is very dry at the top and very moist at the bottom. If the entire layer is lifted to saturation, what will happen to stability of the layer? Explain your answer by discussing the relative amounts of cooling at the top and bottom of the layer.

E. The diagram below shows surface winds blowing around a cyclone in the northern hemisphere. Copy this diagram onto your paper and add vectors showing each of the forces that is acting on the wind. Describe the orientations of your forces relative to the wind.



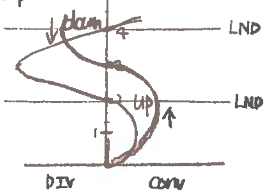
A) Kinematic vert. velocity

Continuity → total DIV = 0 → $\left(\frac{\partial u}{\partial x}+\frac{\partial v}{\partial y}\right)_p+\frac{\partial \omega}{\partial p}=0$

$\rightarrow \int_{\omega_0}^{\omega}\partial\omega=-\int_{P_0}^{P}\left(\frac{\partial u}{\partial x}+\frac{\partial v}{\partial y}\right)_p\partial p$ → consider a pressure weighted vert. average of u, v → <u>, <v>

$\rightarrow \omega-\omega_0=(P_0-P)\left(\frac{\partial\langle u\rangle}{\partial x}+\frac{\partial\langle v\rangle}{\partial y}\right)=(P_0-P)\overline{(\nabla\cdot \underset{\sim}{V})}_p$

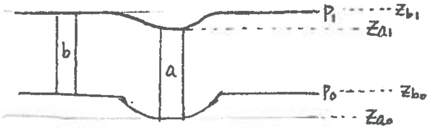
LL conv, coupled with UL div. → upward motion throughout column. However area of div. exceeds conv. below. Hence there should be some downward motion in the UL to feed the horiz. div.



(note: max vert. motions at LND (level of non-divergence, or DIV = 0)

B) The thickness eq or hypsometric eq is $\Delta Z=\frac{R\bar{T}}{g}\ln\left(\frac{P_0}{P}\right)$ ← derive!

So a warm core (a) has a greater thickness between pressure sfcs than the surrounding area (b) for it to be relatively warm. (see diagram)



Therefore for the thickness at the core (a) to be greater than the thickness at the edge (b), the cyclone intensity (as indicated by it pressure depression on a height sfc) must weeken

$$d\Phi=gdz=-\alpha dp=-\frac{RT}{p}dp=-RTd\ln p$$

$$\Phi(z_2)-\Phi(z_1)=R\int_{p_2}^{p_1}Td\ln p$$

$$Z\equiv\Phi(z)/g_0$$

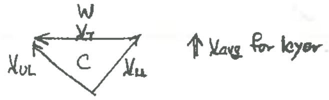
$$Z_T\equiv Z_2-Z_1=\frac{R}{g_0}\int_{p_2}^{p_1}Td\ln p$$

C) Consider the thermal wind eq. $\frac{\partial \underset{\sim}{V}_g}{\partial \ln p}=-\frac{R}{f}(\mathbb{k}\times\nabla_p T)$

Integrating from P_0(LL) → P_1(UL)

$\rightarrow \underset{\sim}{V}_T=\underset{\sim}{V}_{UL}-\underset{\sim}{V}_{LL}=-\frac{R}{f}\int_{LL}^{UL}(\mathbb{k}\times\nabla T)\,d\ln p$ → thermal wind has cold air to left.

So if wind changes CCW with height, there is average Cold advection in the layer



↑ $\underset{\sim}{V}_{avg}$ for layer

D) equivalent potential temp ≡ $\theta_e=\theta\exp\left(\frac{Lq_s}{c_pT}\right)$ As $\frac{Lq_s}{c_pT}\geq 0\rightarrow\theta_e\geq\theta$

θ_e is used to account for the effects of moisture in instability. As a moist parcel rises, it will eventually become saturated and a further rises will lead to condensation → parcel releases heat (latent heat of condensation) If the env. lapse rate > moist adiabatic lapse rate, the parcel will become buoyant and rise. This is called conditional instability.

So $\frac{\partial\theta_e}{\partial z}\begin{cases}<0 & \rightarrow \text{conditional unstable}\\ =0 & \rightarrow \text{saturated neutral}\\ >0 & \rightarrow \text{absolute stable}\end{cases}$

If we lift our layer, the moist bottom will warm more (due to latent heat of condensation) → the layer becomes more unstable ✓

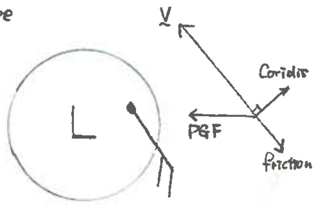
E) Near the sfc, scale analysis indicates three forces important in determining wind.

① Coriolis force: acts to right of wind

② friction: acts opposite of wind direction

decel. wind → reduce Coriolis ($C_o \propto |\underset{\sim}{V}|$)

③ PGF: acts along pressure gradient



In this case, friction has disrupted geostrophic balance

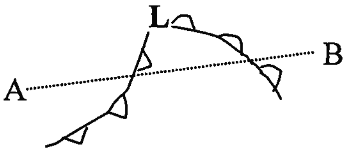
★ Synoptic met.

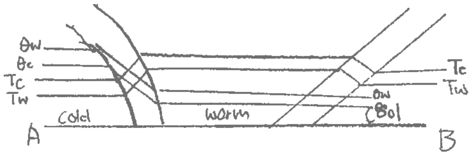
MET4450: Fuelberg (30 minutes)

- The question below mainly is for Mr. someone, both of whom have taken MET 4500 from me. However, since it is an assortment of material from the senior level synoptic course, I believe any graduate student should be able to answer it. I suggest that it be considered for all of the candidates. Completion should require approximately 30 min.

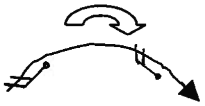
Answer each of the following:

a) Draw a cross section through the wave cyclone shown below. Show two isotherms (labeled T_c and T_w) and two lines of constant potential temperature (labeled θ_c and θ_w). All four isolines should intersect both fronts.





b) You are using a geostrophic wind scale in the flow below at 500 mb. Are the geostrophic wind speeds greater or less than observed values? Based on the observed speeds, assume the geostrophic scale indicates a contour spacing of 0.5 in. To account for the curvature, should you make the contours closer together, farther apart, or make no change?

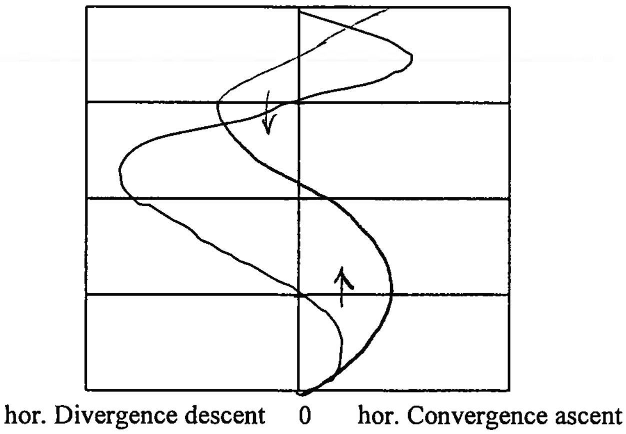


c) The equations for purely divergent flow are given by

$$u = 1/2div.x \qquad v = 1/2div.y$$

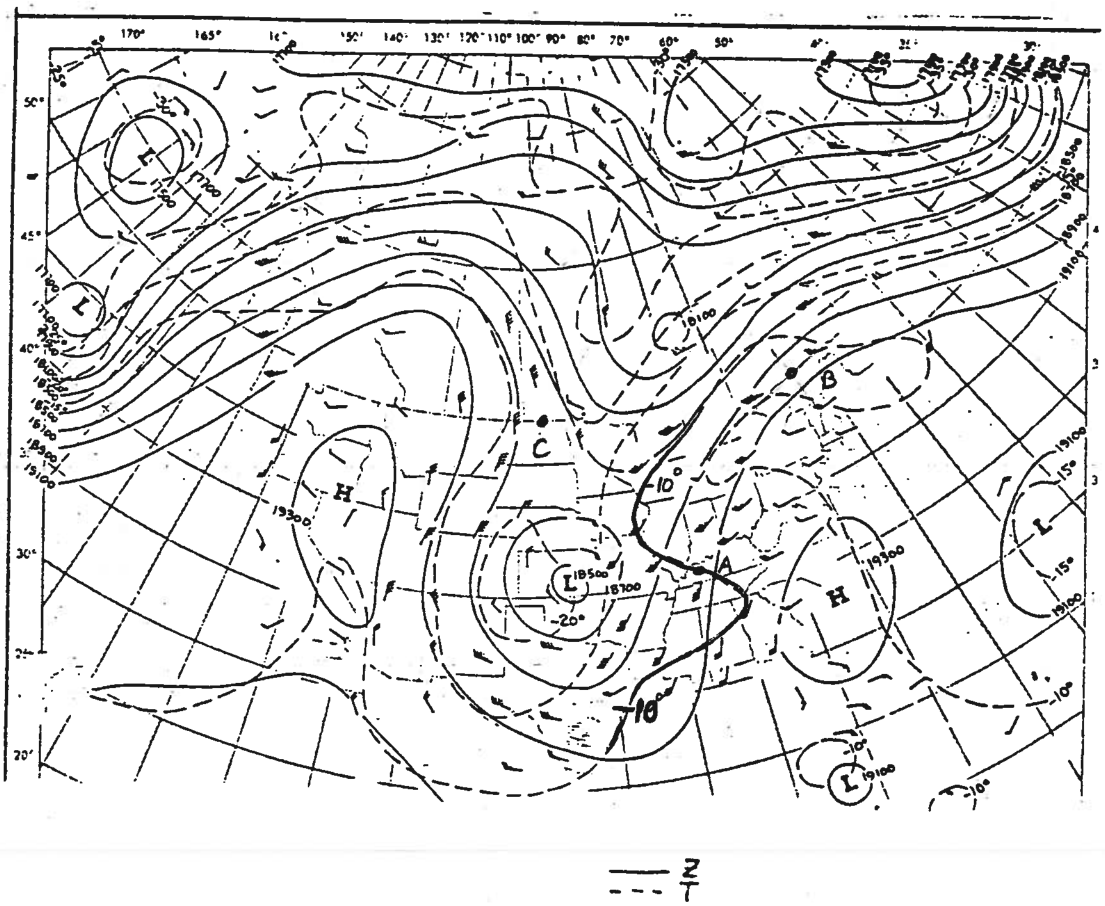
Derive an equation for the streamlines. Show all of the little mathematical steps. Then sketch the streamline pattern.

d) Superimpose on the diagram below a qualitative profile of the corresponding vertical motion using the kinematic approach. Do this very carefully, keeping up with the areas.

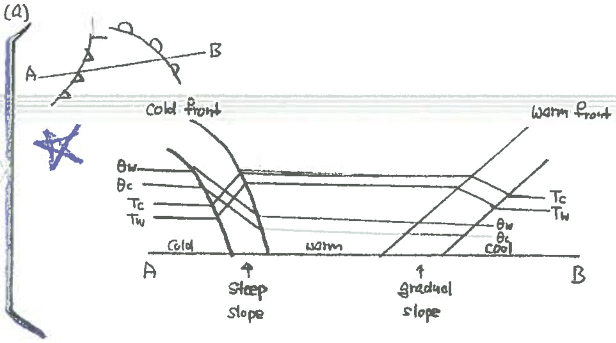


e) Answer the following questions based on the map below. No calculations should be done. Your answers should be based solely on inspection of the map. Explain your reasoning for each answer.

- Is [✓ cold] or warm advection occurring over Kentucky (point a)?
- Does stretching lead to positive or negative divergence at point b located just north of New York?
- Does spreading lead to positive or negative divergence at point c located at the Canada-North Dakota border?
- Does shear lead to positive or negative vorticity at point b?
- Does curvature lead to positive or negative vorticity at point b?



√ (a)



(b)

For gradient wind, $\frac{V^2}{R} + fV = fV_g \longrightarrow \frac{V_g}{V} = 1 + \frac{V}{Rf}$

So as V_g & V defined >0, $R<0$ for anticyclonic $\longrightarrow \frac{V_g}{V} < 1 \longrightarrow V_g < V$

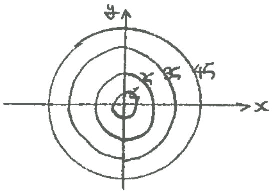
So you should space your isobars <u>further</u> than you would for geostrophic balance.

√ (c)

Helmholtz theorem $\Rightarrow \underset{\sim}{V} = \underset{\sim}{V}_\psi + \underset{\sim}{V}_\chi$, $\underset{\sim}{V}_\chi$ = Vdivergent

$u_\chi = \frac{1}{2}(\nabla\cdot\underset{\sim}{V})x$, $v_\chi = \frac{1}{2}(\nabla\cdot\underset{\sim}{V})y$

$\chi = -\delta\left(\frac{x^2+y^2}{4}\right)$ $\psi = \zeta\left(\frac{x^2+y^2}{4}\right)$



→ Purely divergent flow typically associated with the velocity potential χ

→ $\text{div} = \nabla\cdot\underset{\sim}{V} = \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y}$ ⟶ purely divergent flow is irrotational

$-\frac{\partial\chi}{\partial x} = u = \frac{1}{2}\text{div}\, x$

$-\frac{\partial\chi}{\partial y} = v = \frac{1}{2}\text{div}\, y$

$\nabla^2\chi = -\delta$

$\chi = -\delta\left(\frac{x^2+y^2}{4}\right)$, $\psi = \zeta\left(\frac{x^2+y^2}{4}\right)$

$\begin{pmatrix} u = -\frac{1}{2}\zeta y \\ v = \frac{1}{2}\zeta x \end{pmatrix} \longrightarrow \nabla^2\psi = \zeta$

⟶ wrong! See other notes!

(d) See previous question.

* Synoptic met.

4450

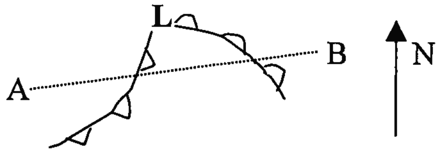
MET?: Fuelberg (?) → two figures needed!

- These questions are meant to test your knowledge of several basic principles of synoptic meteorology. This type of material is covered in our senior level synoptic classes. Answers do not require elaborate mathematical formulations – just simple explanations that demonstrate your basic understanding.

✓ a) What is a front?

Sketch a cross section along axis A-B through the wave cyclone shown below in the surface analysis. On your cross section, show the front as well as a few isotherms and isentropes. Indicate the relatively colder/warmer isoline values.

Assume that the surface analysis is for an oceanic region, i.e., there is no variation in surface elevation. Superimpose on the schematic the 700 mb locations of the low, the cold front, and the warm front. Describe briefly the major points that your sketch is reveling.



✓ b) Define potential temperature, giving its characteristics, and explaining why it is important in synoptic meteorology. Also define equivalent potential temperature, giving its characteristics, and explaining why it is important in synoptic meteorology. How do values compare to those of potential temperature?

✓ c) Sketch three temperature/dew point soundings on your own schematic renditions of Skew T-Log P diagrams (or another common type of thermodynamic diagram) representing the following types of temperature inversion situations:

a) frontal
b) subsidence
c) surface based radiation

Be sure on each diagram to show a sample isotherm, dry adiabat, and saturation mixing ratio line.

d) Answer the following questions based on the attached map. No calculations should be done. Your answers should be based solely on inspection of the map and your meteorological knowledge.

Does horizontal shear lead to cyclonic or anticyclonic vorticity at Point A?
Does curvature lead to cyclonic or anticyclonic vorticity at Point C?
Does horizontal shear lead to cyclonic or anticyclonic vorticity at Point B?
Is there confluence or difluence at Pont B?
Do you expect upward or downward vertical motion at Point B? (explain)

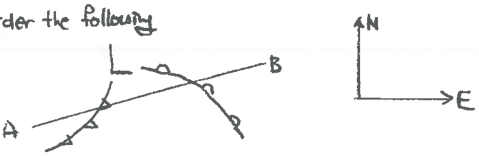
e) Answer the following questions based on the attached sounding. No calculations are necessary – answer on the basis of inspection.

What kind of horizontal temperature advection (warm or cold) is occurring between the surface and 800 mb?
Based on the parcel criteria, what is the stability of the 800-700 mb layer?
How will equivalent potential temperature vary with height increasing altitude in 900-700 mb layer?
Based on the convective stability criteria, what is the stability of the 900-700 mb layer?

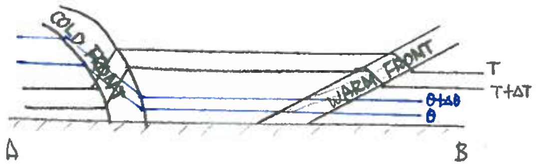
How will potential temperature vary with increasing altitude in the 600-500 mb layer?
If the 600-700 mb layer is lifted <u>without any saturation occurring</u>, how will its stability change?
If the 600-700 mb layer experiences horizontal <u>divergence without any saturation occurring</u>, how will its stability change?

a) A front?

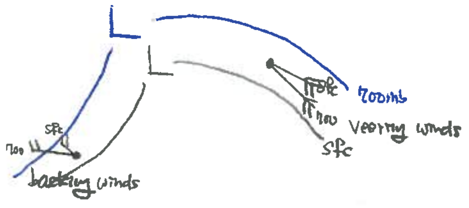
- A front is often defined as an elongated zone of strong temperature gradient and relatively large static stability + cyclonic vorticity. It may also be defined in terms of density or moisture, or other derived variables like θ or θe. "Strong" means at least an order of magnitude larger than the typical synoptic scale value of 10°C/1000 km. A zone whose length is roughly half an order of magnitude or more larger than its width is elongated.
- Atmospheric fronts may be defined as sloping zones of pronounced transition in the thermal and wind fields. They are characterized by a combination of relatively large horizontal temperature gradients, static stability, absolute vorticity (horizontal wind shear) + vertical wind shear. When depicted on quasi-horizontal surfaces, fronts appear as long, narrow features in which the along front scale is typically an order of magnitude greater than the cross-front scale (1000-2000 km compared with 100-200 km). Fronts are shallow phenomena with depths typically 1-2 km. Because of this geometrical configuration, horizontal variations in the cross front direction greatly exceed those in the along front direction
- Consider the following



A typical cross section through AB would look like
(see fig 3.12 p128 of Wallace + Hobbs)



(see fig 3.14 p130 of Wallace + Hobbs)



→ System tilts towards colder air to the NW
→ 1:200 slope of warm front vs 1:50 slope of cold front means that at 700mb warm front lies much further north of sfc warm front than is the case for the cold front.
→ stronger winds at 700 mb than sfc
→ trough associated with sfc low is less distinct + displaced upstream to NW
→ warm frontal rain commonly coincides with location of upper level frontal zone.

b) θe vs θ

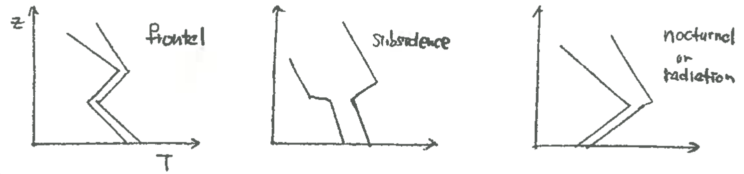
θ ≡ (dry adiabatic) potential temperature
↳ temperature an air parcel would have if brought to a reference pressure level (generally 1000mb) dry adiabatically.

$$\theta = T\left(\frac{P_o}{P}\right)^{R/C_p} \qquad P_o = 1000 \text{ mb (usually)}$$

θe ≡ equivalent potential temperature $\theta_e = \theta \exp\left(\frac{L q_s}{C_p T}\right)$
↳ lift a parcel dry adiabatically to saturation and then continue along the moist adiabat until all the moisture is condensed out of the parcel. Assume the latent heat of condensation was used to heat the air parcel. Bring the air parcel dry adiabatically down to a pressure level (usually 1000mb). The parcel temperature at this point is the equivalent potential temperature

θ is conserved for dry adiabatic processes
θe is conserved in moist adiabatic processes (in addition to dry adiabatic processes
↳ by design θe accounts for the buoyant effects of water vapor in a parcel
since $\frac{L q_s}{C_p T} \geq 0$ and $\theta_e = \theta \exp\left(\frac{L q_s}{C_p T}\right)$ we have $\theta_e \geq \theta$

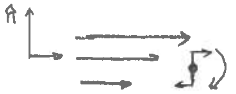
c) Inversions



* Note for vorticity + divergence

In natural coordinates

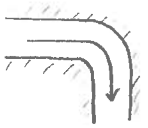
$$\zeta = \underset{\uparrow \text{ curvature vorticity}}{\frac{V}{R_s}} - \underset{\uparrow \text{ shear vorticity}}{\frac{\partial V}{\partial n}} \qquad \text{and} \qquad \text{div} = \frac{\partial V}{\partial s} + V\frac{\partial \theta_b}{\partial n}$$



$\frac{\partial V}{\partial n} > 0$ so $-\frac{\partial V}{\partial n} < 0$

no curvature vorticity as $R_s \rightarrow \infty$

negative (anticyclonic) shear vorticity

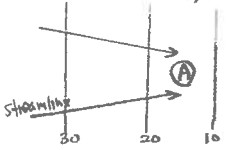


$R_s < 0$ for anticyclonic curvature

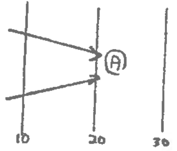
$V = \frac{ds}{dt} \geq 0$ so

$\zeta < 0 \rightarrow$ negative curvature vorticity

Confluence at Ⓐ



Confluence at Ⓐ but speed divergence



* various aspects of synoptic met.

4450

MET? (?): Fuelberg (?)

- Question

i. Make a list of procedures or rules that you would used to locate a cold front or a warm front on a surface map. Explain why each rule is appropriate.

ii. You are given a 500 mb map of the middle latitudes. The data are very sparse. Your job is to analyze height contours. You understand the concept of geostrophic wind – contour spacing is closer in regions of strong winds. However, your map has regions of cyclonic flow and anticyclonic flow. How do you adjust geostrophic contour spacings into those that would account for curved flow? Specifically, will the spacing be closer or farther apart in a) cyclonic and b) anticyclonic flow? Explain your answers by discussing the forces and concepts associated with the geostrophic and gradient winds.

iii. If winds change direction in a clockwise manner with increasing height, what type of temperature advection is occurring? Explain your answer with diagrams and a discussion

warm adv.

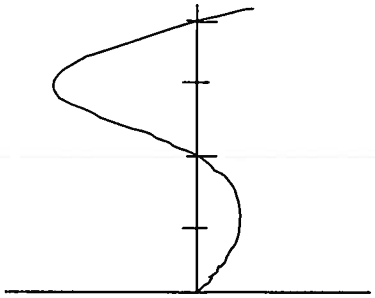
iv. A layer of air is dry at the top and moist at the bottom. Its equivalent potential temperature decreases in the vertical. If the entire layer is lifted to saturation, what will happen to the lapse rate of the layer. Explain your answer by discussing the relative amounts of cooling at the top and bottom of the layer.

v. The isentropic coordinate system is becoming increasingly popular. What are the advantages and disadvantages of this system? What parameters usually area analyzed on isentropic surfaces?

→ See Barclow's question

vi. Forecasts often say that positive vorticity advection at 500 mb suggests ascending vertical motion. Using the vorticity equation or some other appropriate equation, explain why this is true.

vii. The diagram below contains a profile of horizontal divergence. Superimpose a sketch of the vertical motion profile. You will not know actual values (number) of the ascent ~~of~~ or descent.



I. Procedure to locate a warm/cold front on a sfc map

(a) Isotherm analysis → locate regions of strong temperature gradients
fronts on cold air side of ∇T

(b) Isodrosotherm analysis → locate moist + dry air.

(c) locate areas of clouds + precipitation → nephanalysis

(d) Isallobaric analysis → pressure tendency suggests movement via pressure rises + falls. rapid pressure rise behind cold front

(e) Isobaric analysis → frontal boundaries lay in pressure trough.

(f) examine windshift, perform streamline analysis.

(g) clean up analysis → proper kinking of isobars around front.

II.

Make use of the gradient wind eq. in the form

$$\frac{V_g}{V} = 1 + \frac{V}{fR} \quad \left(\text{from } \frac{V^2}{R} + fV = fV_g\right)$$

where
$V \equiv$ gradient wind ≥ 0 by definition
$V_g \equiv$ geostrophic wind
$f \equiv$ Coriolis parameter
$R \equiv$ radius of curvature $\begin{cases} R>0 & \text{for cyclonic flow} \\ R<0 & \text{for anticyclonic flow} \end{cases}$

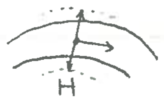
Note that in regions of cyclonic flow $V_g > V$ since $R>0$

since Ce works with Co implied P is smaller

We observe V which is less then V_g. However, we use the gradient value, V, with our geostrophic wind scale. We get a weaker pressure gradient than exists in reality.

The actual wind is near gradient wind balance. As $V_g > V$ in regions of cyclonic curvature we say the flow is subgeostrophic. Given V we would reduce the spacing between Φ lines to get $V_g > V$.

In regions of anticyclonic flow $R<0$, $V_g < V$. The wind is super geostrophic

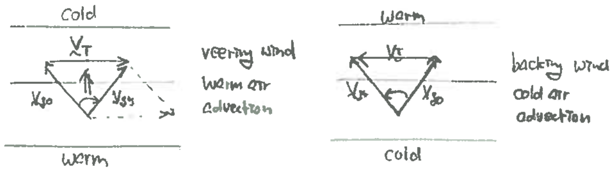


The gradient wind is stronger than the geostrophic wind. Thus you should relax the geopotential gradient to less than that implied by the gradient wind.

III.

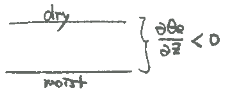
Thermal wind → $\frac{\partial \mathbf{V}_g}{\partial \ln p} = -\frac{R}{f}\mathbb{k} \times \nabla_p T$

Winds backing with increasing height → cold air advection
" veering " " " → warm air advection



IV. Layer instability

Consider the following layer of air



$\frac{\partial \theta_e}{\partial z} < 0$

Now lift this layer to saturation
What happens to the lapse rate?

The bottom of the layer is moist. As we lift it, it reaches saturation more quickly than the drier layer on top. After reaching saturation the temp. of the bottom increases along the moist adiabat. The dry top of the layer takes longer to reach saturation while moving up the dry adiabat to saturation. When both the bottom + top of the layer are saturated the bottom will have travelled along the moist adiabat more than the top layer. Hence the bottom will be warmer than the top. An unstable lapse rate is the result.

V. look at other notes

VI. PVA at 500mb → rising motion, why?

Consider the vorticity eq in the form

$$\frac{d}{dt}(\zeta + f) = -(\zeta + f)(\nabla \cdot \mathbf{V})$$

$\uparrow$
time rate change in absolute vorticity following the motion $= -\begin{pmatrix}\text{absolute}\\\text{vorticity}\end{pmatrix}$(horizontal divergence)

Expand the LHS and rewrite as

$$\frac{\partial}{\partial t}(\zeta + f) = -\mathbf{V} \cdot \nabla(\zeta + f) - \omega \frac{\partial}{\partial p}(\zeta + f) - (\zeta + f)(\nabla \cdot \mathbf{V})$$

Mid-latitude scale analysis suggests that we can neglect the vertical advection of vorticity so that

$$\frac{\partial}{\partial t}(\zeta + f) = -\mathbf{V} \cdot \nabla(\zeta + f) - (\zeta + f)(\nabla \cdot \mathbf{V})$$

Assume $\frac{\partial}{\partial t}$ is small compared to horizontal advection $(-\mathbf{V} \cdot \nabla(\zeta + f))$ and the divergence term $(-(\zeta + f)(\nabla \cdot \mathbf{V}))$. Positive vorticity advection, $-\mathbf{V} \cdot \nabla(\zeta + f) > 0$, implies $-(\zeta + f)(\nabla \cdot \mathbf{V}) < 0$. As $(\zeta + f) > 0$ generally we must have $-\nabla \cdot \mathbf{V} < 0$ or $\nabla \cdot \mathbf{V} > 0$ $\Rightarrow$ horizontal divergence. Now via mass continuity upper level divergence in a column must be balanced by low level convergence in a hydrostatic atm. Low level convergence along with upper level divergence imply rising motion.

$\Rightarrow$ PVA at 500mb implies rising motion.

- 500mb PVA implies $\zeta_{500} \uparrow$. In quasi-geostrophic dynamics $\zeta_g = \frac{1}{f_0}\nabla^2\Phi$. Therefore $\zeta_{g500} \uparrow$ implies $\nabla^2\Phi \uparrow$ or $\Phi \downarrow$. Falling heights can occur with the adiabatic cooling associated with rising motion.

VII.

Recall continuity eq in pressure coordinates

$$\frac{\partial \omega}{\partial p} = -\nabla \cdot \mathbf{V} \longrightarrow \text{at LND we have } \frac{\partial \omega}{\partial p} = 0 \text{ as } \omega \text{ is max or min}$$

Integrate from P_s to P

$$\int_{P_s}^{P} \frac{\partial \omega}{\partial p} dp = -\int_{P_s}^{P} \nabla \cdot \mathbf{V} \, dp$$

$$\omega(p) - \cancel{\omega(p_s)} = +\overline{(\nabla \cdot \mathbf{V})}\,\Delta p$$

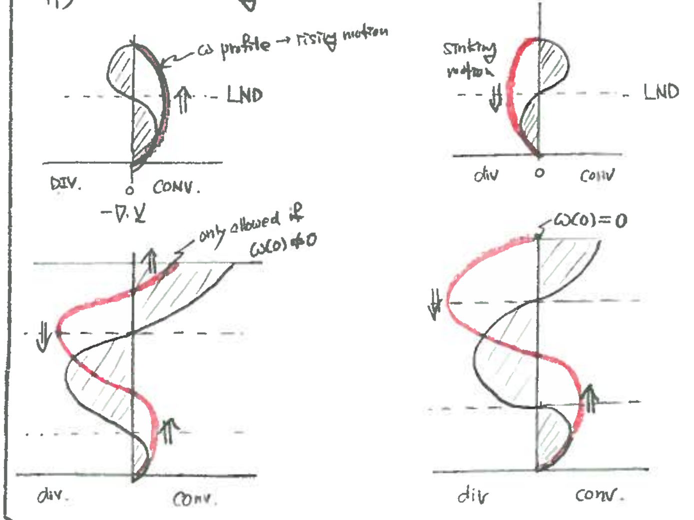
↳ vertically integrated horizontal divergence.

as a BC,
assume $\omega(p_s) = 0 = \omega(0)$ → these B.C.'s imply area of convergence = area of divergence

So

$\omega(p) = \overline{(\nabla \cdot \mathbf{V})}\,\Delta p \longrightarrow$ net convergence implies $\omega < 0 \longrightarrow$ rising motion
net divergence implies $\omega > 0 \longrightarrow$ sinking motion

Apply this to the following $\nabla \cdot \mathbf{V}$ profile



* Convection related questions

✓ MET5511C: Fuelberg (45-60 minutes) *

- Question

A. Areas of convection are known to affect their larger scale environments. Describe these effects on fields and/or vertical profiles of temperature, geopotential height, winds (including divergence/convergence).

✓B. What is a "mesoscale convective complex"? How is it defined, and how does it differ from other types of convective storms? → see Kloesel's note

✓C. What are atmospheric gravity waves (describe them)? How are they thought to trigger thunderstorm formation?

A.

* 1) Thermal perturbations associated with convection ⇒ low-level height falls, upper-level height rises.
   - Low-level cooling from precip. (evaporative cooling)
   - upper-level warming from condensation

2) Horiz. DIV at upper levels ; Horiz. CONV at low-levels of convection
   → amplification of vertical velocity field.

3) After convection decays, Φ + ζ perturbations remain
   → feed back to large scale environment.

B. others

C. see next question.

A

① low-level height fall + upper level height rise due to thermal perturbations associated with convection → low level cooling due to precipitation
(upper level heating from condensation of vapor

② horizontal convergence at low levels and divergence in upper levels
→ through continuity of mass these pattern imply a large amplification of vertical velocity

③ after convection decays Φ + ζ perturbations remain. Thus areas of convection can feed back to the large scale environment.

B. see Kloesel's exam

*gravity waves

MET? : Fuelberg (30-45 minutes) — gravity waves.

✓ • Describe atmospheric gravity waves by giving as many characteristics as you can (wavelength, amplitude, etc.). In what kinds of atmospheric conditions of stability, wind shear, etc. do gravity waves tend to occur? Describe how gravity waves might help to initiate thunderstorm development. ⟶ also see Kloesel's note.

See Holton sec. 7.3.2

*

* Gravity waves are more appropriately called buoyancy waves, and result from buoyancy oscillation. They are transverse waves (propagate perpendicular to the direction of oscillation). Due to sloping of height/pressure sfcs, they can propagate vertically as well as horiz. They can be formed by thermodynamic and/or dynamic triggers:

1) external dynamic forcing (i.e. orographic lifting)
2) shear instability → $Ri = \dfrac{\frac{g}{\theta}\left(\frac{\partial\theta}{\partial z}\right)}{\left(\frac{\partial u}{\partial z}\right)^2} < \dfrac{1}{4}$ (critical Ri for onset of Kelvin-Helmholtz waves)

For gravity waves to exist + propagate, the atmosphere must be stable. The greater the stability, the higher the freq. of gravity waves. Also the atmosphere must have large vertical wind shear!

Effects of gravity waves include:

a) spectral energy transfer
b) vert./horiz. transport of momentum + energy
c) generation of CAT
d) triggers of instability in PBL → development of severe WX.

* The top of the PBL is ~ stable, so the waves can exist there. These waves can ease changes to DIV field in the PBL, which can increase low-level CONV in an area → enhanced upward motion → Potential trigger for convection.

* microburst/downburst

✓ MET? : Fuelberg (30-45 minutes)

- Describe microbursts and downbursts by giving typical sizes, lifetimes, wind speeds, and any other information that shows your understanding. How do microbursts/downbursts differ from tornadoes? How have microbursts led to several major airline disasters? → also see Kloesel's note

Sol)

parameter	Downburst	Microburst
Width	~4-20 km	<4 km
life time	5-30 mins	5-10 mins
speed	≲ 60 m/s (horiz.)	≲ 75 m/s (horiz.)

Basically downbursts/microbursts are same physical phenomene.
Microbursts are basically small but intense downbursts.
Downbursts are highly divergent winds, producing strong downward motion.

This downward motion has led to aircraft to sudden lose altitude + crash.

Downbursts vs. Tornadoes : Whereas downbursts are highly divergent, tornadoes are highly convergent. Also tornadoes are the product of strong vertical shear tipping strong horiz. vort. into the vertical.

*
Tornadoes are highly convergent, swirling winds affecting a relatively narrow path.
Downbursts are highly divergent straight line winds of damage causing intensity
Straight line wind – nondivergent straight line winds of damage causing strength seen just behind an advancing gust front.

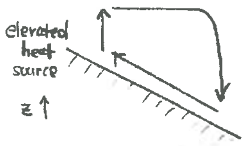
* sea/land breeze + slope/valley circulation

MET? : Fuelberg (45 minutes)

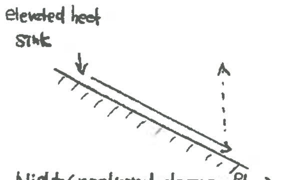
- Mesoscale circulations can easily dominate synoptic scale flows. In dispersion calculations, for example, transport due to mesoscale processes can be quite different from that due to assumed large scale flow. It is important to be aware of the characteristics of mesoscale circulations.

1. Give a complete three-dimensional description of the sea/land breeze circulation at day and at night. Don't be shy about giving information here. I suspect it will require several handwritten pages to complete. → see Kbesol's note
2. Give a similarly complete description of slope/valley circulation at day and at night.

2. Slope/valley circulation

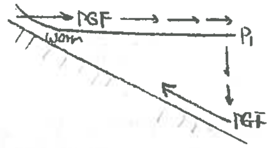


DAY (Day time upslope flow)
deeper flow (layer 100s of meters)



Night (nocturnal drainage flow)
Shallow flow (10's of meters)

vertical components ~ 3-6 m/s



Consider daytime :
The air near the slope heats
→ warmer than env. air at that level. Invoke hydrostatic → hypsometric → at some elevated level, the pressure must increase over that at some level away from slope → PGF away from slope. As air (a greater pressure) moves away from slope, this creates a pressure gradient force at sfc toward slope (from hydrostatic), which gives an upslope component
→ Circulation!

Day : These circulations can provide trigger for convection in mountain areas.
Night : These circulations → drainage of cooler, more stable area in valleys
→ implications on pollution + fog.

2. Give a detailed discussion of the 3D structure + characteristics of Slope-valley circulations at day and night.

In a region with irregular terrain, local wind patterns can develop because of the differential heating between the ground surface and the free atmosphere at the same elevation some distance away. A larger diurnal temperature variation usually occurs at the ground, so that during the day the higher terrain becomes an elevated heat source, whereas at night it is an elevated heat sink. Two categories of winds are usually recognized ⇒ slope flow + valley winds. These types of flow are easiest to observe when the prevailing large scale flow is light.

Slope flow refers to (1) cool, dense air flowing down elevated terrain at night.
(2) warm, less dense air moving toward higher elevations during the de

We refer to flow (1) as nocturnal drainage flow. Flow (2) is referred to as daytime upslope flow. You will also hear the nocturnal drainage flow called a katabatic wind ; anabatic for the daytime upslope flow.

Valley winds are up- and down-valley circulations that develop from along-valley horizontal pressure gradients in one segment of a valley. Such gradients occu because of the input, by the slope flow, of air having a different temperature structure than the air in adjacent segments.

- Slope-valley circulations are of great importance in determining local weather wh major synoptic disturbances are not around. Unlike sea/land breezes, these circulations are not easy to characterize. The coupling of slope/valley circulations is more involved than in sea/land breezes. Observations suggest that for light synoptic forcing the daytime upslope flows are ~ 3 to 6 m/s and the depth of the layer involved is several hundred meters or more. Night time valley flows have the same speed but are much shallower → on the order of several tens of meters deep The depth of the nocturnal drainage flow increases linearly in the downslope directi

- Mechanism

The circulation theorem of Bjerknes provides the basic explanation for the sea/land breeze. Similar arguments apply to the thermally induced slope/valley circulation. For simplicit only one slope is considered and the initial condition is assumed to be cloudless.

⟨Daytime⟩

→ absorption of solar radiation by the sloping ground leads to warming of the near sfc air.
→ the air near the sloping surface becomes warmer than the air in the free atmosphere at the same height above sea level.
→ Invoke the hydrostatic principle + consider the isobars to be initially horizontal. Now the vertical pressure gradient will be greater in the cooler free air than in the warmer air near the slope. This means that at a given height within the influence of slope heating, pressure will be higher than at a point further away from the slope This horizontal pressure gradient causes air to move away from the slope + thus generates higher pressure than had been previously experienced at a lower level - say over an adjacent plain. Consequently a low level, horizontal presun gradient from plain to slope is created + an upslope wind results. The circulation in the vertical plane is equivalent to that in a sea breeze.

- At night the mechanism and the circulation are reversed. Surface cooling causes an upper level pressure gradient towards the slope and a cold air drainage down and away from the slope.

※ Trivia : valley/slope circulations appear to be more efficient than sea/land breezes in that they require less heat input to generate a circulation of comparable size but greater intensity.

- Convective cloudiness associated with slope-valley flows

The upward velocities associated with daytime upslope flows and frequent interaction with synoptic flow provide in many mountaneous areas a triggering mechanism for cloud formation and afternoon showers. Both modeling + observational studies indicate the importance of daytime, thermally forced flows to local weather

- Nocturnal valley flows.

During the nocturnal period when the sky is clear + the synoptic flow is light, the basin of a valley is characterized by the accumulation of relatively cold, stable and stagnant air. Vertical mixing by turbulence is often negligable at night so that the cold air accumulates in the bottom of valleys + basins. This has important implications with regards to

(a) air quality in drainage basins ↗ pollution
↘ fog formation

(b) breakup of nocturnal inversion following sunrise + development of daytime PBL. Failure to break the inversion (perhaps due to snow covered sfc) can lead to a prolonged accumulation of pollution.

(c) Frost/freeze threat to agriculture in drainage basin

- Low level jet

Various theories for the development of the LLJ. One mechanism views formation of the LLJ as a response of the atmospheric boundary layer to the diurnal thermal forcing over sloping terrain. The horizontal ∇T that is created by the diurnal heating and cooling can result in a LLJ by creating a large thermal wind component. Modeling studies indicate that the diurnal boundary heating over the sloping terrain of the Great Plains is directly associated with the formation of the LLJ. Maddox (1983) suggested a linkage between development of MCCs and the warm advection involved with the LLJ over the Great Plains

- Mountaineous Coastal area Circulations

The presence of mountains next to water bodies result in a circulation that blends sea/land breezes with mountain/valley flows. The cooler air advected inland by the sea breeze can reduce the thermal forcing for upslope flow. Orographic precipitation may be enhanced along the sea breeze. If the sea breeze flows over a terrain barrier, it can experience significant warming as it descends down the other side of the barrier.

1/

* land/sea breeze

MET5511 (Mesometeorology): Fuelberg (30 minutes)

- Land/sea breeze are important mesoscale circulations near sea coasts. Describe them in detail, being sure to include the following items in your discussion:

- mechanisms producing them
- a cross section normal to the coast during the day showing the circulations
- a cross section normal to the coast during the nigh showing the circulations
- a description of how the wind direction at a point slightly inland varies during a 24 h period
- a description of the sea breeze front
- effects of the following on sea breezes:
  - → larger scale gradient wind
  - → atmospheric stability
  - → topography
  - → shape of the coast line
- any other material that you believe is appropriate.

* land/sea breeze (differential heatings)

MET? (Mesometeorology): Fuelberg (?)

- Question

a) A number of mesoscale circulations are due to differential heating at the surface. List as many of these as you can. For each circulation, very briefly describe the type of differential heating that serves as the forcing mechanism.

b) Provide a detailed description of land/sea breezes. As part of your discussion, give information about wind directions and speeds and how these vary during a 24-hour period, the sea breeze front and its influence on local weather, inland penetrations, vertical structure, and any other information that is appropriate.

c) How can the shape of a coast line affect the sea breeze?

* atmospheric stability

MET? (Synoptic Meteorology): Fuelberg (45 minutes)

- The following question has several parts dealing with atmospheric stability.

✓ a) You have the "sounding" for a station that is plotted on a thermodynamic diagram (such as the Skew T). Describe how you can VISUALLY determine the parcel stability of any portion of the sounding. Please describe your procedure for both saturated and unsaturated regions.

✓ b) An unsaturated layer of the atmosphere is 100 mb thick. Its lapse rate is 7 deg C/km. This entire layer is lifted 200 mb; however, none of the layer becomes saturated during the ascent. How will the lapse rate change during the ascent (i.e., will it become larger, smaller, or stay the same)? EXPLAIN why this occurs.

c) A different layer of the atmosphere also is 100 mb thick, and its lapse rate also is 7 deg C/km. This layer merely experiences horizontal divergence. None of the layer becomes saturated during the process. How will the lapse rate change? EXPLAIN why this occurs.

d) A layer of the atmosphere has a lapse rate of 7 deg C/km. It is dry at the top of the layer, but moist at the bottom of the layer. The layer is lifted until all portions become saturated. How will the lapse rate change as a result of the ascent? EXPLAIN why this occurs.

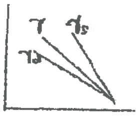
e) How can we use the vertical derivative of equivalent potential temperature in describing part b) above?

* Note for stability

For dry adiabatic

$\frac{\partial\theta}{\partial z} < 0$ unstable
$\frac{\partial\theta}{\partial z} = 0$ neutral
$\frac{\partial\theta}{\partial z} > 0$ stable

} Same for moist parcel if you replace θ with θ_e



In a temperature inversion, $\theta\uparrow$ with height ($\frac{\partial\theta}{\partial z} > 0$). In a superadiabatic layer $\gamma > \gamma_d \Rightarrow$ potential temperature decreases with height ($\frac{\partial\theta}{\partial z} < 0$)

- Parcel method

$\gamma = \gamma_d \longrightarrow \frac{\partial\theta}{\partial z} = 0 \leftarrow$ dry neutral
$\gamma = \gamma_s \longrightarrow \frac{\partial\theta_e}{\partial z} = 0 \leftarrow$ moist neutral

$\gamma < \gamma_s$ absolutely stable
$\gamma_s < \gamma < \gamma_d$ conditionally stable
$\gamma > \gamma_d$ absolutely unstable

- Layer method

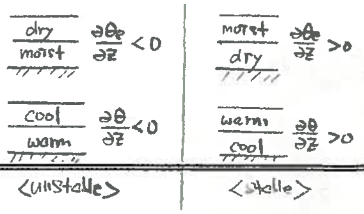
Consider $\gamma_d - \gamma = \text{const} * p'$

case ①: stable $\gamma < \gamma_d$
constants are positive
lift the layer
p' decreases
γ increases
∴ less stable

case ②: superadiabatic $\gamma > \gamma_d$
constants are negative
lift the layer p' decreases
γ gets smaller
the layer is not as unstable

- Convective or potential stability criteria

$\frac{\partial\theta_e}{\partial z} > 0$ stable
$\frac{\partial\theta_e}{\partial z} = 0$ neutral
$\frac{\partial\theta_e}{\partial z} < 0$ unstable



☆ Lifting entails a decrease in stability
→ lift layer → θ surface expand → decrease stability

☆ divergence entails increasing stability
convergence " decreasing "

* multicell + supercell storms

MET5511: Fuelberg (30 minutes)

- Question

a) Draw schematic diagrams and use them to explain the important features of a multicell severe storm complex.

b) Draw schematic diagrams and use them to explain the important features of a supercell storm.

→ see Kloesel's note

* multicell + supercell storms + microbursts

MET5511C: Fuelberg (45 minutes)

- Question

✓ A. Fully describe the structure and evolution of multicell storm complexes. This is an open ended question that allows you to tell all that you know.

see Kloesel's note

✓ B. Fully describe the structure and evolution of supercell storms. Once again, give all the information that you know.

C. Describe microbursts fully, contrasting them with tornadoes where appropriate.

→ see the previous question.

* dry lines

√

MET5511C: Fuelberg (1 hour) *

- Give a full discussion of 1) dry lines, 2) "lids" (or "caps") in a thermodynamic sounding, and 3) thunderstorm development along dry lines. You should cover at least the following topics in your answer. Any other material is always helpful.

    How a dry line is located on a surface analysis.
    How a dry line evolves during a 24 h period and WHY this occurs.
    Sketch an east-west cross section through a dry line and EXPLAIN features of that cross section.
    Sketch a sounding on the dry side of the dry line and another on its moist side. DISCUSS important aspects of each sounding
    What is the origin of air on each side of the dry line.
    Explain the stability characteristics on each side of the dry line.
    Discuss the lid strength index and how it compares to more common types of stability indices.
    Finally, discuss the concepts that a forecaster should consider regarding thunderstorm formation along the dry line, i.e., play like you are forecaster considering this area.

* A case study of a dry line : MAY 11-12, 1991

1. Introduction

A dryline, as generally defined, is a region of sharp dewpoints gradients which is often observed in the central United States during spring and summer. Drylines have a three dimensional structure which is quite evident from looking at upper level data. Furthermore, diurnal variations in drylines are also observed. Like fronts, drylines are synoptic scale in one dimension and mesoscale in two dimensions. Drylines, as atmospheric boundaries, are distinguishable from fronts in that the temperatures need not increase or decrease across them. Indeed, at varying times both of these temperature distributions may be observed.

The dryline being studied in this paper is a classic case. All of the above characteristics are present and will be discussed at some length.

2. Surface characteristics

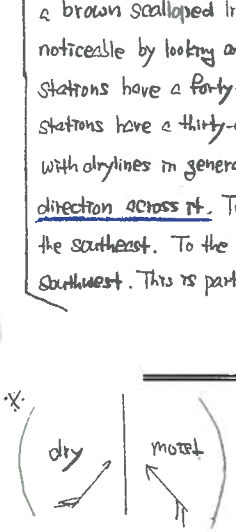
As stated previously, the identifier of a dryline is the strong gradient of dewpoint across it. Fig. 1 shows a station model plot of the south central United States at 0Z on May 11, 1991. The dryline is marked by a brown scalloped line. The dewpoint gradient across the dryline is clearly noticeable by looking at station pairs. In southwest Texas neighboring stations have a forty-two degree °F difference. Further north, cross-dryline stations have a thirty-degree °F dewpoint difference. Although not present with drylines in general, this dryline also has a marked shift in wind direction across it. To the east, in the moist air, winds are generally from the southeast. To the west, in the dry air, winds are generally from the southwest. This is particularly evident in southeast New Mexico.

3. Vertical characteristics

Soundings from stations in the area of the dryline are useful in understanding its vertical structure. Fig. 2 shows a cross section from Tuscon, Arizona to Longview, Texas for 0Z on May 11, 1991. Contours are equivalent potential temperature at intervals of four Kelvin. In this case, the equivalent potential temperature gradient is as useful as the dewpoint gradient for locating the dryline. The dryline rises vertically to approximately 750 mb and then curves over the moist air. The vertical wind profile is also quite striking. While the wind direction is nearly constant with height to the west of the surface dryline, there is marked veering with height at the stations to the east of the surface dryline. Fig. 3 shows the 850 mb map for the same time. As expected from the cross section, the dryline is still quite evident at this level by simply looking at the dewpoint depressions. On the other hand, if the cross section is accurate, we would expect the dryline to not be in evidence at 700 mb. Fig. 4 shows the 700 mb map for the same time. The air at this level is almost uniformly dry and shows little evidence of any moisture gradient. Thus, this dryline is confined to the troposphere below 700 mb. Fig. 5 is a north-south cross section from Del Rio. Texas to Dodge City. Kansas for 0Z May 12, 1991. Contours are equivalent potential temperature at intervals of one Kelvin. The equivalent potential temperature gradient between 850 mb and 700 mb clearly defines the dryline. This figure reveals the planar characteristics of the dryline. The dryline extends into and out of the plane of this cross section. In a 3D approach it is perhaps a misnomer to label this boundary a dry "line".

Fig. 6 shows the sounding at Del Rio, Texas for the same time as above. At this time Del Rio is to the east of the surface dryline in the moist air. The dryline is evident above the station between 700 mb and 850 mb. The air above is much drier than the air below. Furthermore, the dryline is marked by a stable layer or cap. Fig. 7 shows the sounding at North Platte, Nebraska at the same time. This station is also east of the surface dryline at this time. The dryline is again clearly marked above the station by a sharp decrease in dewpoint and an inversion layer. It can be

※



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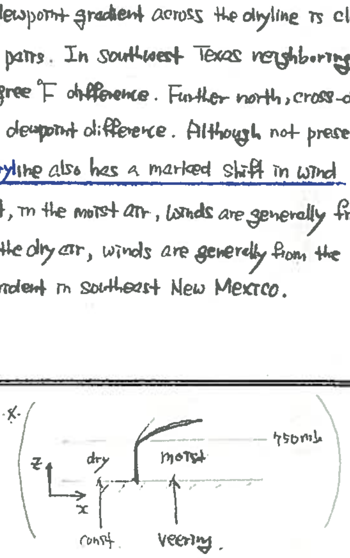
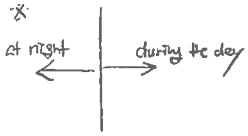


concluded therefore that *the air east of the surface dryline is capped by a dry air inversion.

Fig. 8 shows the sounding at El Paso, Texas for the same time. This station is to the west of the surface dryline. The air is dry throughout the atmosphere. Fig. 9 shows the sounding at Albuquerque, New Mexico for the same time, also west of the surface dryline The sounding is quite similar to El Paso's, dry throughout. Both soundings have a weak inversion layer and a jump in the dewpoint sounding around 600mb. This is perhaps evidence of an upper level type II front which extends upward to the west

4. Temporal variations

The dryline undergoes diurnal changes Figures 1, 10, 11, 12 and 13 form a time series of surface charts at six hourly intervals starting at 0Z May 11, 1991. Between 0Z and 6Z on May 11, the dryline is seen to move westward. Between 12Z and 18Z on May 11, the dryline is seen to move eastward This is symptomatic of the motion of drylines in general. At night, dry lines will tend to move westward. During the day, drylines will tend to move eastward. Fig 14 shows the meteorogram trace for Roswell, New Mexico between 0Z May 11 and 0Z May 12. The top panel shows the temperature and dewpoint traces in degrees Fahrenheit. The dewpoint rises sharpely as the dryline initially moves past the station westward. The wind also shifts to the southeast direction, previously described as associated with the moist air. Then, in the afternoon, the dryline moves back eastward past the station. The dewpoint falls sharply and the temperature rises, corresponding to the warmer, drier air to the west of the dryline. The same behavior is evident in fig. 15, the meteorogram trace for Cannon Air Force Base, New Mexico. The dryline moves westward past the station during the night, and then eastward past the station during the afternoon.

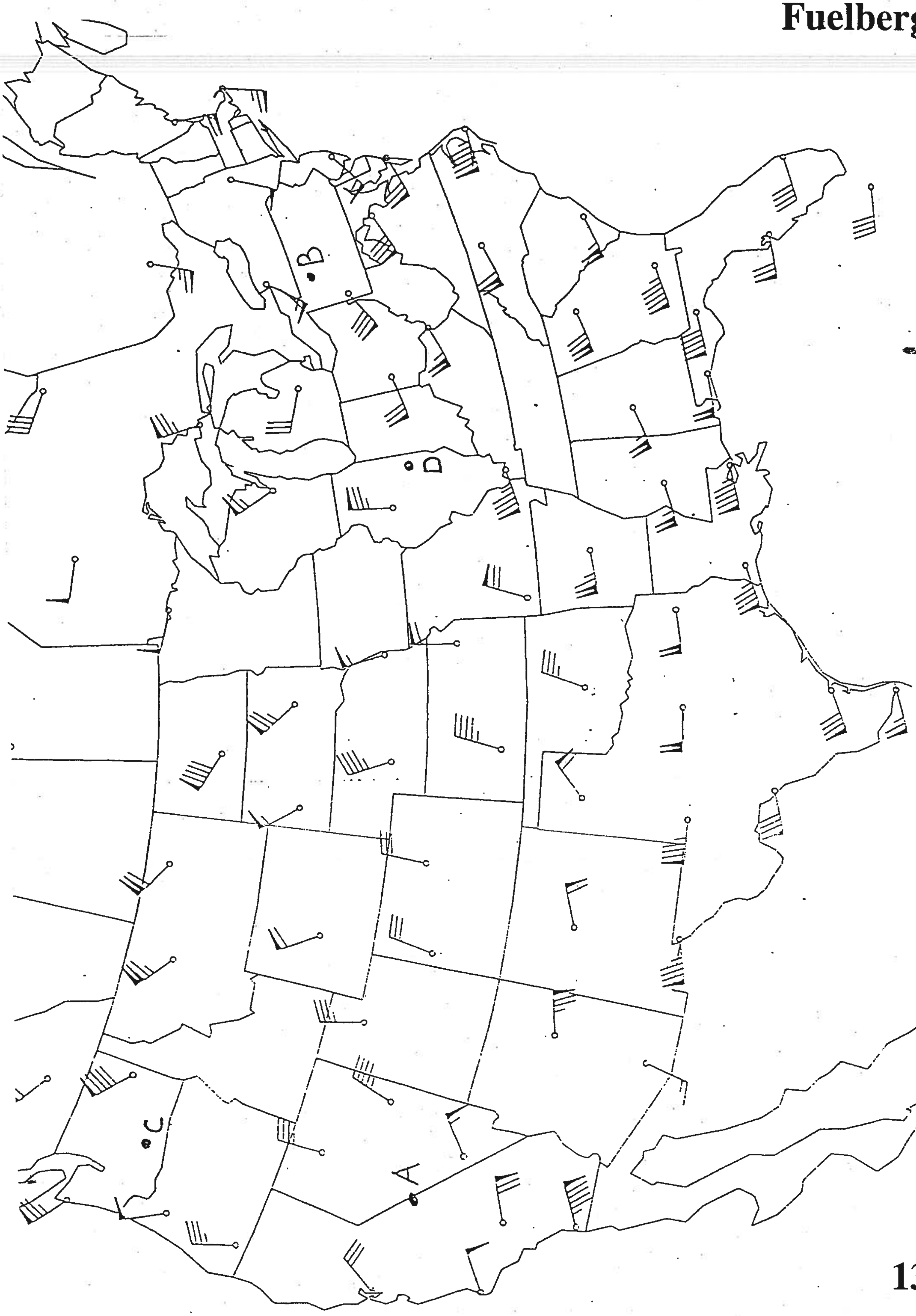


This motion is connected to the reversal of the temperature gradient across the dryline between night and day. Fig. 1 shows the surface chart at 0Z, or late in the day. The temperatures to the west of the dryline are warmer than the temp. to the east of the dryline. Because the air is drier, it can heat up more rapidly. Also, clouds will tend to form in the moist air to the east of the dryline which further limits the temp. in the moist air. In fact, most of the stations to the east of the dryline are reporting clouds.

Fig. 11 is the surface chart for 12Z, or early morning. The opposite temp. gradient is evident here. Temperatures west of the dryline are cooler than temperatures east of the dryline. The same characteristics which allowed the dry air to be warmer during the day, now make it cooler during the night. The drier air radiates faster than the moist air east of the dryline. Also, the greater amount of cloudiness in the moist air keeps the temperature up at night. Once again, many of the stations to the east of the dryline are reporting clouds.

5. Conclusion

The discussion above revealed the 4D characteristics of the dryline on May 11-12, 1991. The dryline fit well into the widely accepted theories concerning the structure and time evolution of drylines. Furthermore, the dryline was very well defined on all analyses. It is probable that this dryline is a stronger than average dryline.



Plot of 300 mb Wind barbs for 0Z 28 OCT 93

* understanding of WX maps.

4450

MET?(Synoptic meteorology): Fuelberg(45 min) : ***Two figures needed!!***

- Use the enclosed 300 mb data plot to answer the following questions

1. Does shear lead to positive or negative vorticity at point A? Discuss this.
2. Does curvature lead to positive or negative vorticity at point A? Discuss this.
3. Repeat questions 1 and 2 for point B.
4. Does stretching lead to positive or negative divergence at point C? Discuss this.
5. Does spreading lead to positive or negative divergence at point C? Discuss this.
6. How will geostrophic wind speeds compare to gradient wind speeds at point A? How will they compare at point B? Explain.

Answer the following questions based on the enclosed thermodynamic diagram.

7. Is cold or warm temperature advection occurring between the surface and 850 mb? Explain this.
8. How does equivalent potential temperature vary with height between 850 and 700 mb (pos. or neg.)? What is the convective stability category of this layer? Explain.
9. How does potential temperature vary with height between 500 and 400 mb? Explain.
10. In what direction do cold core surface low pressure centers tilt with height? Explain your answer using the concept of thickness.

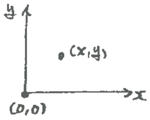
★ Basic kinematics of the horizontal wind

(Fuelberg)

We will deal with the horizontal wind $V_H = u\mathbf{i} + v\mathbf{j}$

We use the Cartesian coordinate system

We wish to describe the flow at point (x, y) which is close to the origin



The flow at (0,0) is given by u_0, v_0

We expand u + v as 2D Taylor series

$$u = u_0 + \frac{\partial u}{\partial x}\Big|_0 x + \frac{\partial u}{\partial y}\Big|_0 y + \frac{1}{2}\frac{\partial^2 u}{\partial x^2}\Big|_0 x^2 + \frac{1}{2}\frac{\partial^2 u}{\partial y^2}\Big|_0 y^2 + \frac{\partial^2 u}{\partial x \partial y}\Big|_0 xy + \cdots$$

$$v = v_0 + \frac{\partial v}{\partial x}\Big|_0 x + \frac{\partial v}{\partial y}\Big|_0 y + \frac{1}{2}\frac{\partial^2 v}{\partial x^2}\Big|_0 x^2 + \frac{1}{2}\frac{\partial^2 v}{\partial y^2}\Big|_0 y^2 + \frac{\partial^2 v}{\partial x \partial y}\Big|_0 xy + \cdots$$

To a good approx

$$\begin{cases} u \simeq u_0 + \frac{\partial u}{\partial x}\Big|_0 x + \frac{\partial u}{\partial y}\Big|_0 y \\ v \simeq v_0 + \frac{\partial v}{\partial x}\Big|_0 x + \frac{\partial v}{\partial y}\Big|_0 y \end{cases}$$

One can show that the above may be rewritten as

$$\bigstar\ u \approx \underbrace{u_0}_{ⓐ} - \underbrace{\frac{1}{2}\left(\frac{\partial v}{\partial x} - \frac{\partial u}{\partial y}\right)_0 y}_{ⓑ} + \underbrace{\frac{1}{2}\left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y}\right)_0 x}_{ⓒ} + \underbrace{\frac{1}{2}\left(\frac{\partial u}{\partial x} - \frac{\partial v}{\partial y}\right)_0 x}_{ⓓ} + \underbrace{\frac{1}{2}\left(\frac{\partial v}{\partial x} + \frac{\partial u}{\partial y}\right)_0 y}_{ⓔ}$$

and

$$\bigstar\ v \approx \underbrace{v_0}_{ⓐ} + \underbrace{\frac{1}{2}\left(\frac{\partial v}{\partial x} - \frac{\partial u}{\partial y}\right)_0 x}_{ⓑ} + \underbrace{\frac{1}{2}\left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y}\right)_0 y}_{ⓒ} - \underbrace{\frac{1}{2}\left(\frac{\partial u}{\partial x} - \frac{\partial v}{\partial y}\right)_0 y}_{ⓓ} + \underbrace{\frac{1}{2}\left(\frac{\partial v}{\partial x} + \frac{\partial u}{\partial y}\right)_0 x}_{ⓔ}$$

where

ⓐ u_0, v_0 → translation → no change in area, shape, orientation

ⓑ $\frac{\partial v}{\partial x} - \frac{\partial u}{\partial y}$ → vorticity → no change in area + shape but orientation changes

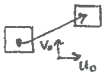
ⓒ $\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y}$ → divergence → no change in shape + orientation but area changes

ⓓ $\frac{\partial u}{\partial x} - \frac{\partial v}{\partial y}$ → stretching deformation → change shape.

ⓔ $\frac{\partial v}{\partial x} + \frac{\partial u}{\partial y}$ → shearing deformation

· eg. for streamlines

→ translation ⇒ $u = u_0$; $v = v_0$



→ vorticity ⇒ $u = -\frac{1}{2}\zeta y$; $v = \frac{1}{2}\zeta x$

$$\zeta = -\frac{\partial V}{\partial n} + V\frac{\partial \theta_b}{\partial s}$$

→ curvature term

→ shear term

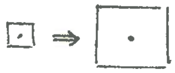
→ divergence ⇒ $u = \frac{1}{2}(\nabla \cdot \underset{\sim}{V})x$; $v = \frac{1}{2}(\nabla \cdot \underset{\sim}{V})y$

$$\text{div} = \frac{\partial V}{\partial s} + V\frac{\partial \theta_b}{\partial n}$$

→ confluence/difluence term

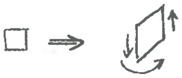
→ turning of streamlines

→ speed convergence / divergence



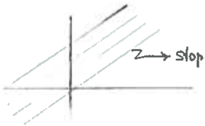
→ Stretching deformation : $\frac{\partial V}{\partial s} - V\frac{\partial \theta_b}{\partial n}$

→ shearing deformation : $\frac{\partial V}{\partial n} + V\frac{\partial \theta_b}{\partial s}$



✓ Transition

$$\frac{dy}{dx} = \frac{v}{u} = \frac{v_0}{u_0} \quad \text{so} \quad dy = \frac{v_0}{u_0}dx \Rightarrow y = \frac{v_0}{u_0}x + \text{const}$$



→ slope $\frac{v_0}{u_0}$

→ Speed is same everyw (no isotachs)

✓ vorticity

$$\frac{dy}{dx} = \frac{v}{u} = \frac{\frac{1}{2}\zeta x}{-\frac{1}{2}\zeta y} \rightarrow \frac{dy}{dx} = -\frac{x}{y} \rightarrow x\,dx + y\,dy = 0 \rightarrow \boxed{x^2 + y^2 = k}$$

circle

→ Isotachs

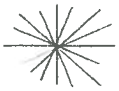
$|\underset{\sim}{V}| = \frac{1}{2}\zeta(x^2+y^2)^{1/2}$ — concentric circles

✓ Divergence

$$\frac{dy}{dx} = \frac{v}{u} = \frac{\frac{1}{2}(\nabla\cdot\underset{\sim}{V})y}{\frac{1}{2}(\nabla\cdot\underset{\sim}{V})x} \rightarrow dy = \frac{y}{x}dx \rightarrow d\ln y = d\ln x \rightarrow \ln y = \ln x + \text{const}$$

$$\rightarrow \ln\left(\frac{y}{x}\right) = \text{const} \rightarrow \boxed{y = (\text{const})x}$$

← straight line



→ Isotachs

$|\underset{\sim}{V}| = \frac{1}{2}(\nabla\cdot\underset{\sim}{V})(x^2+y^2)^{1/2}$

✓ Stretching deformation

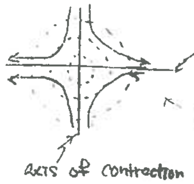
$$\frac{dy}{dx} = \frac{v}{u} = \frac{\frac{1}{2}D_1 y}{-\frac{1}{2}D_1 x}$$

$$\frac{dy}{dx} = -\frac{y}{x}$$

$$d\ln x + d\ln y = 0$$

$$\boxed{xy = k}$$

hyperbola



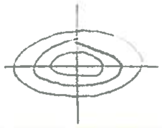
✓ Shearing deformation

$$\frac{dy}{dx} = \frac{v}{u} = \frac{\frac{1}{2}D_2 x}{\frac{1}{2}D_2 y}$$

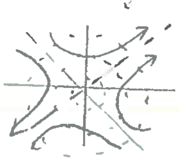
$$y\,dy - x\,dx = 0$$

$$\boxed{x^2 - y^2 = k}$$

→ ellipse (?)



→ hyperbolas (but the axes of dilitation + contraction have been rotated 45°)



MET 4500
Fuelberg

- Answer the following questions in the spaces provided or on the enclosed analyses. Give complete answers that demonstrate your knowledge of the material. Assume that the grader (Fuelberg) knows very little – you must explain everything. If key points of an answer are omitted, he assumes that you didn't know them! When diagrams are requested, you should explain those diagrams.

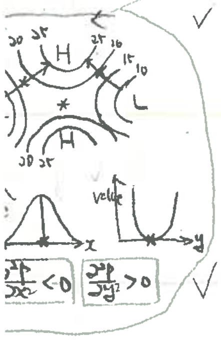
✓ 1. How does a warm core anticyclone vary in intensity with increasing altitude? Based your answer on the hypsometric equation, draw appropriate diagrams, and EXPLAIN those diagrams.

✓ 2. What is truncation error? How is it involved in finite differencing? When will centered, forward, and backward differencing give the same answers for a first derivative? When will you get different answer? A sketch of each situation would be helpful in your discussion.

✓ 3. How can you use surface isobars to get a feel for the speed of a front? Sketch a front/isobar configuration for a rapid moving front. Sketch a second configuration for a much slower moving front. Explain.

✓ 4. Draw a graph of pressure vs. time for a 24 hour period. No pressure changes other than the semidiurnal pressure wave are occurring. Label the times of your maxima and minima.

   Sketch isobars for a col (→ or saddle point) in surface pressure. You should have about 8 isobars in all. Use a base of 1000 mb and an interval of 4 mb. Label the high and low regions.



✓ 5. The geostrophic wind at a lower level is from the west at 10 kt. The thermal wind from that level to a higher level is from the south at 10 kt. What are the geostrophic wind direction(deg) and wind speed(kt) at the top of the layer? Show your work below.

   The geostrophic wind at a lower level is from the south at 10 kt. The geostrophic wind at an upper level is from the east at 20 kt. What are the direction (deg) and speed (kt) of the thermal wind in this layer? Show your work below.

✓ 6. Use the enclosed analyses of temperature and height contours. Calculate temperature advection at the station in eastern Kentucky where the wind is from 235^0 at 35kt. Use centered differences with $\delta x = \delta y = 2^o$ lat. Express your answer in degC/day. Show all of your work (unit conversion… everything below).

✓ 7. Use the enclosed temperature/height analysis to answer the following questions.
   a. At point B draw a vector indicating the orientation of the temperature gradient.
   b. Give the signs of $\delta T / \delta x$ and $\delta T / \delta y$ at these points: Do this by inspection. Do no calculations. WAA CAA
   c. Use the "cherries and berries" approach to give temperature advection at all intersections the map.
   d. Where on the map is cold air temp. advection the greatest in magnitude? Print "max" (large red letters) at that area.

✓8. Answer the following questions based on the enclosed analysis of thickness between 1000 and 500 mb. There may be several correct answers. But you only need to give one of them. Notations on the analysis should be in large red letters.

   a. Place an "X" at the location where geostrophic winds vary most in the layer.
   b. Pace an "Y" at the location where geostrophic winds vary least in the layer. Pick a location between two thickness lines – not at the edge of the analysis.
   c. Place a "Z" at the location where the mean virtual temp. in the layer is the coldest.
   d. Draw the orientation of the thermal wind vector at point "F" on the map.
   e. At point T, the wind direction at 1000 mb is from the south at 10 kt, is cold or warm advection occurring in the 1000-500 mb layer?

✓9. Draw a cross section through a cold front. Label the warm and cold areas. Draw two isotherms (solid lines) intersecting the front, labeling them T and $T+\delta T$. Also draw two isolines of potential temperature (dashed) intersecting the front, labeling them θ and $\theta+\delta\theta$.

- Answer the following questions in the spaces provided. Give complete answers that demonstrate your knowledge of the material. Assume that the grader (Fuelberg) knows nothing – you must explain or show everything to hie. If key points of an answer are omitted, he assumes that you don't know them. When diagrams are requested, you also mush EXPLAIN those diagrams.

✓1. The sketch below shows a frontal wave and an associated dry line. Likely air masses are mT, cT, and mP. Label these air masses in their proper locations on the diagram. If this analysis were for 5 AM, would the coldest air likely be located east or west of the dry line? Explain your answer below.

✓2. Define frontogenesis in terms of what we see on a typical constant pressure analysis. List the four processes that cause frontogenesis. Each is represented by a term of an equation. 1. 2. 3. 4.
Describe ONE of these four processes in detail, using a sketch as part of your answer. Your example should describe frontogenesis – NOT frontolysis.

✓3. Draw a cross section through <u>a cold type occluded front</u>. No isotherms or theta lines are needed. Label the various regions cold, coldest, and warm. Show the direction of motion with an arrow.
Where in the United States are cold occluded fronts typically found?
Which locations have maximum cyclogenesis in the vicinity of the contiguous United Sates?

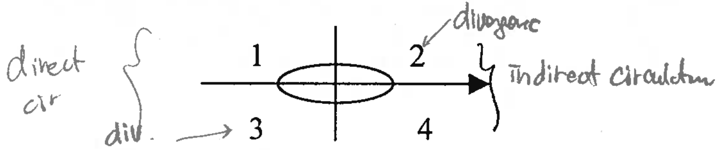
✓4. Where does cyclogenesis tend to occur with respect to an upper level trough/ridge pattern? Be very specific.
What kind of vorticity advection occurs there? Is there upper level divergence or convergence in that region?

Sketch a low and associated cold, warm, and occluded fronts. Superimpose the typical cloud pattern, clearly showing the "dry slot". Then, superimpose <u>numerous</u> arrows over the whole region showing upward and downward motion. Finally, explain the cause for the dry slot in terms of vorticity advection and temperature advection.

5. Sketch streamlines AND isotachs for a situation of pure positive shearing deformation. Label the axes of contraction and dilatation.

   Sketch isogons at 30 deg. Intervals for a case of pure positive vorticity that is centered at the point below. Label your isogons.

6. The diagram below is a jet streak at 300 mb. I have divided it into four regions that are labeled 1 – 4. Fill in the blanks below, noting that an answer may (or may not) consist of more than one quadrant.



The entrance region? __1, 3_____

Area containing rising motion? __2, 3_____

Area containg upper level divergence? __2, 3_____

What is the typical forward motion of jet streaks? ___30___kts → |V| = 150 kt.

7. Answer the following question using the enclosed wind plot. Use centered differences with $dS = dn = 2$ deg lat ($2ds = 2dn = 4$ deg lat). Let 2 deg lat = 1 cm. Give your answers in 10^{-5} s^{-1}.

   Calculate the divergence due to stretching at point A. Use natural coordinates, showing your axes on the diagram (in red), and ALL of your input data AND calculations below.

   Calculate the divergence due to spreading at point A. Use natural coordinates, showing your axes on the diagram (in red), and ALL of your input data AND calculations below.

8. Answer the following questions based on inspection (no calculations).

   Does shear lead to positive or negative vorticity at point B_________

   Does curvature lead to positive or negative vorticity at point B_________

   Do we expect the gradient wind to be stronger or weaker than the geostrophic widn at point B________

   Does stretching lead to convergence or divergence at point D_________

   Should the actual contour spacing at point E be tighter or looser than the plotted wind suggest? _________

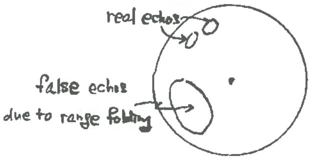
9. Use the box (inflow/outflow) procedure to estimate horizontal divergence at point C. Use the winds at Buffalo, NY and Chatham, MA in the calculations. Show your box on the diagram (in red) and all of your input data and calculations below. Express your answer in 10^{-5} s^{-1}.

MET5511C – Mid-term : Fuelberg.

1. Describe the concepts of range folding and velocity folding in words. How would each appear on a radar display?

Sol)

Velocity folding is a type of aliasing which occurs with higher radial velocities. For example, if the atmospheric radial velocity is 60 kts for a particular place, the radar may not receive that value correctly. (appear slower than actual) The value will be lower than the actual. Velocity folding occurs due to the PRF, or pulse repetition frequency, which is usually about $1000 s^{-1}$. We can minimize velocity folding by choosing a higher PRF. This will serve to increase our maximum unambiguous velocity, but at the same time, our maximum unambiguous range will decrease. On a radar display, we may therefore see areas of precipitation which are not really there due to this range folding.



To see if we are being affected by this phenomena, we can tweak the PRF a little in order to see if our suspect area of radar echos is moving or not.

2. List and very briefly describe the factors which adversely affect precipitation estimates from radar.

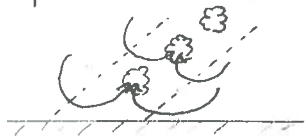
Sol)
- not all precipitation droplets (different drop size distributions?) will be similar for different types of storms.
- not all precipitation droplets are liquid
- Because the radar beam gains altitude with increasing distance from the radar site, some of the precipitation at the lower levels of the atmosphere will not be sampled properly.
- The volume of the beam will not always be filled completely by precipitation droplets.
- Incorrect calibration.

3. How are the inland penetration and strength of the sea breeze affected by the large scale flow? Refer to Arritt's paper in your discussion, but you don't have to give all of the specifics that he does. How does hydrostatic stability affect the sea breeze? Why does it occur?

Sol)
- Offshore large scale flow : Sea breeze does not penetrate as quick due to opposing flow; sea breeze is stronger with offshore flow due to a strengthening of $\nabla_s T$ → Stronger ∇p ⇒ Stronger convergence at low levels/divergence at upper levels → Stronger vertical motion ⇒ more instability ⇒ possible strong convection.
  The optimal value for the strongest sea breeze is an offshore flow value of around 3 m/s. However, we can still have inland penetration for speeds as high as 11 m/s.
- Onshore large scale flow : farther penetration, but weaker sea breeze circulation as $\nabla_s T$ not as strong.
- Calm large scale flow : Similar to but not as strong as offshore case.
- Hydrostatic stability
  - more unstable conditions ⇒ more vertical mixing ⇒ depth of ∇T is greater (∇T extends through a deep layer) ⇒ this strengthens the entire sea breeze circulation ⇒ Stronger sea breeze.
  - more stable conditions ⇒ less vertical mixing ⇒ depth of ∇T is less ⇒ weaker sea breeze.

4. Describe horizontal convective rolls in words and sketches. Explain how they appear in visible satellite imagery and why this occurs. How might HCRs influence convective development along a sea breeze front?

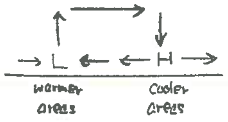
Sol) HCR's are areas of rotating air about a horizontal axis which leads to lines of cumulus parallel to mean flow



They appear on visible imagery as cloud streets or bands oriented parallel to the mean flow.
As the sea breeze moves toward these HCR's, the vertical motions will be enhanced where the air is rising in the above sketch. This could lead to more significant convective development along the sea breeze front.

5. What is an inland sea breeze? What situations might produce one? How does the circulation of an inland sea breeze vary during the course of a daylight period? Sketches might help here.

Sol)
Inland sea breeze : an area where there are ∇T which are stronger than normal. These may arise due to soil type, soil moisture, vegetative cover, cloudy areas vs clear areas, etc.
During daylight, low level flow is from the cooler areas to the warmer areas, just as in coastal sea breeze. Lower pressures occur in the warm location while higher pressures form in cooler areas.



- return flow sets up aloft ; convection may form in area of enhanced convergence vertical motion.
- most intense during late afternoon ; small area of ↑, larger area of ↓
- vertical motion pattern becomes asymmetric late in afternoon.

6. List at least six factors that influence lake effect snow systems (LES). Comment briefly on each factor.

Sol)
- Instability : The deeper the mixed layer ⇒ steeper lapse rate ⇒ more instability ⇒ more favorable for LES
  Also, if $T_{lake} > T_{850}$ by 13°C ⇒ more favorable for LES as lapse rate approaches dry adiabatic.
- fetch : The longer the fetch over a lake → more favorable for LES
- Wind shear : the less the directional shear + the less the speed shear ⇒ more favorable for LES
  optimal : directional shear < 30°
  If directional shear > 60°, probably will not have LES
- upstream moisture : the more upstream moisture ⇒ easier for saturation + condensation to take place.
- presence of upstream lakes : Same effect as upstream moisture.
  If flow passes over an upstream lake, more favorable for LES downwind of second lake.

- large scale forcing : Synoptic scale vorticity ⇒ vertical motions on synoptic scale, enhancing mesoscale vertical motions.

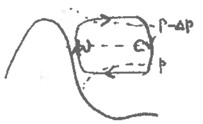
  CAA ⇒ steepening of lapse rate ⇒ more instability.

7. Describe/Distinguish between the following terms :

   anabatic/katabatic flows
   slope circulations
   valley circulations
   anti…flows.

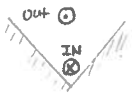
Sol)

- anabatic flow : relatively warm air rising up the side of a mountain during daytime :



warmer air on mountain slope ; cooler air at some altitude over valley. ∇p from high to low pressure due to these ∇T

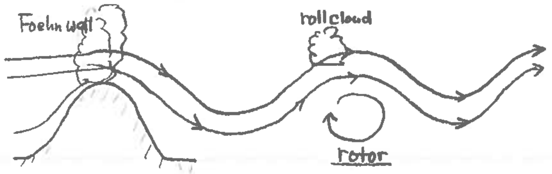
- Katabatic flow : relatively cool air sinking along mountain slope during nighttime (opposite situation from anabatic flow)
- Slope circulation : occurs when air flows down the slope of a mountain.
- Valley circulation : daytime flow in the valley where flow goes toward higher elevations. (a larger scale)
- Anti flows : return flow on the top of slope or valley circulation. (flows in opposite direction)



8. Sketch a cross section through a mountain. Include streamlines that dipict lee waves and a rotor. Also show clouds corresponding to a foehn wall and a roll cloud.

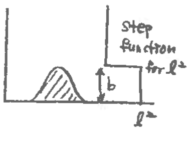
   List several factors that are conducive to lee wave formation

Sol)



Factors conducive for lee waves

- Static stability near mountain top ; stable environment
- Strong horiz. flow
- Wind shear
- Scorer parameter : $\ell^2_{lower} - \ell^2_{upper} > \pi/4b^2$



9. Describe a foehn , Describe a bora

   Several hypotheses for downslope flows have been advanced. Specifically, Briefly describe the "blocking" hypothesis.

   " " " "hydraulic jump" hypothesis

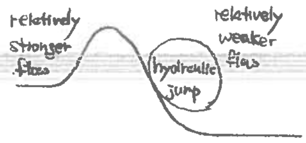
Sol)

foehn : dry, warm downslope wind
bora : dry, cool downslope wind
} → compared to environment.

Blocking hypothesis : mountain acts as a block to horiz. flow

If $h > \pi/\ell$, then mountain can effectively act as a block to low level flow. ℓ : Scorer param. which contains static stability and winds

Hydraulic jump :



As relatively stronger flow goes over mountain, a strong adjustment is needed ⇒ a shock wave, or hydraulic jump occurs

10. Prepare a sketch that looks down on the circulation to the lee of an isolate island. Describe the various features of your sketch and give any other useful information

Sol) → look at Holton's book.

11. There are two major categories of ways by which gravity waves are formed. List these two categories, and give as many specific examples for each as you ca What is the Richardson number? How is it a factor in diagnosing gravity waves

Sol)

- Gravity waves which ride along a external forcing

  These are often due to convection, frontogenesis or cyclogenesis whereby the atmosphere is trying to get back into geostrophic balance after ageostrophic condit
- Shear instability gravity waves, which often dissipate fairly quickly.

  These may be due to vertical motions giving rise to pressure perturbations

  $$R_i = \frac{buoyancy}{shear^2}$$

  If $R_i < 0.5$, favorable for gravity waves.

  Most favorable situation for gravity waves (to last longer) :

  stable layer where wind speed < phase speed of gravity wave

  Above the stable layer ⇒

  a less stable layer where $R_i < 0.25$ and wind speed ≥ phase speed.

12. Sketch a cross section through a gravity wave near the surface. Show surface winds that indicate convergence/divergence at the proper locations. Show areas of rising/sinking air at the proper locations

Sol) skip.

13. Describe/Compare/Contrast the following terms :

    conditional instability, potential instability, convective instability

    Define CAPE and CIN in your own words. How do these relate to vertical moti

Sol)

- Conditional instability : occurs when a parcel's lapse rate is in between the di + moist adiabatic lapse rate. An unsaturated parcel is stable, but if it can become saturated, it will be unstable. This is simply the parcel method No layers are considered.
- Convective or potential instability : These essentially are the same. If θ_e decreases with height and we are saturated, then we are convectively or potentially unstable with respect to the saturated layer.
- CAPE : Convective available potential energy : positive buoyancy area on a sounding between the LFC and equilibrium level (EL). units : J kg

  $W = \sqrt{2 \cdot CAPE}$
- CIN : Convective inhibition : the cap or area of negative buoyancy in the lower layers of the atmosphere below the LFC. Once this cap is weakened, the CAPE may be realized.

  $W = \sqrt{2 \cdot CIN}$ (to reach the LFC)

Synoptic scale vertical motions will not weaken the cap sufficiently. Instead, synoptic scale vertical motions can lead to mesoscale vertical motions which may weaken the cap. If this occurs, we can realize the CAPE and strong/severe thunderstorms may occur.

14. Describe in your <u>own words</u> the concept of inertial instability.

An equation describing inertial instability is

$$\frac{dV}{dt} = -f\left(f-\frac{\partial U_g}{\partial y}\right)dy$$

Let dy be positive, i.e., a northerly displacement. Now, assume a case of <u>anticyclonic shear</u>. What possible stability or instability can occur?

Sol)

- <u>Inertial instability</u> : If a parcel is moving horizontally, its momentum must be conserved. We examine u, f, M_P, M_A, and $\frac{\partial V}{\partial y}$ to see if the parcel is inertially stable (whereby it returns to its initial position) or if the parcel is inertially unstable (whereby it will keep on moving in the direction it was moving)
- Anticyclonic shear

② ⟶ If the change in f from ① to ② is greater than $\partial U_g/\partial y$

① → $\Rightarrow$ inertially stable as $dv/dt < 0$

If the U_g increases more than f $\Rightarrow$ inertially ustable as $\frac{dv}{dt} > 0$

Neutral if shear term $= f$

15. <u>Compare</u>/<u>contrast</u> <u>conditional symmetric instability</u> with hydrostatic and inertial instability.

How do you <u>diagnose</u> CSI using cross sections?

Sol)

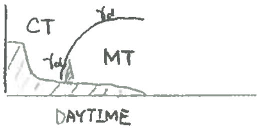
One can have CSI without having hydrostatic instability or inertial instability. This is because CSI is a slantwise instability, while hydrostatic instability is vertical and inertial instability is horizontal.

CSI is diagnosed by looking at cross sections of θ_e and M_g (absolute geostrophic momentum). If the M_g lines are more horizontal (less sloped) than the θ_e lines, one may expect CSI. If $\partial\theta_e/\partial z|_{M_g} < 0$ or if $\partial M_g/\partial x|_{\theta_e} < 0$, then we may expect CSI

16. How do dry lines <u>move</u> during a typical 24h period?

<u>Explain</u> why this occurs. A cross section might help.

Sol) Dry lines move E during day as sun mixes out capped, moist layer near the surface. The dry line moves W during the night as winds decouple, mixing ends, and easterly wind don't die off as quickly within the moist side.



⟹ too brief

MET5511C – Final exam : Fuelberg (1997)

1. Describe the <u>nocturnal low level jet</u> stream. Do this through a numbered list of its characteristics. For each numbered item, provide a contrast with the polar jet stream. You should give at least 4-5 "meaty" characteristics

Sol)

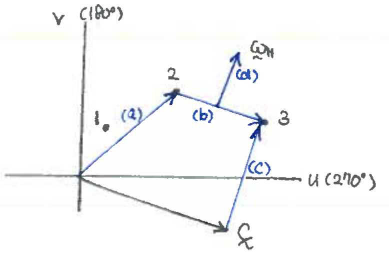
Nocturnal Low level jet	Polar jet
(a) found in low levels, usually below 700mb. At night, can be found as low as 500 m AGL (above ground level)	(a) found in the upper levels, usually near 300mb, but may be slightly higher.
(b) Strong diurnal variation: Subgeostrophic during day; supergeostrophic (e.g. $V_{obs} = 2V_{geos}$) during night	(b) no diurnal variation; ageostrophic circulation may be associated with jet streaks (entrance & exit regions)
(c) found in midwest/Plains	(c) found at mid-high latitudes, practically everywhere around the earth
(d) can be accountable for rapid transport of moisture + airmass destabilization; severe thunderstorms may be associated with this jet.	(d) no moisture transport at low levels, but may be associated with severe thunderstorms when jet is strong (jet streaks). Ageostrophic wind increase at low levels associated with upper jet streak
(e) Core of jet can be found near top of radiation inversion	(e) Core of jet found in area of strong upper level confluence → PJ is associat with tropopause inversion

2. The hodograph below has numbers showing the winds at 1 km intervals. The vector <u>C</u> indicates the motion of a storm cell.

a. Draw a vector indicating the <u>ground relative</u> wind at 2km.
b. " " " " " <u>Shear</u> between 2-3km
c " " " " " <u>Storm relative</u> flow at 3km
d. " " " " " <u>horizontal vorticity</u> between 2-3km

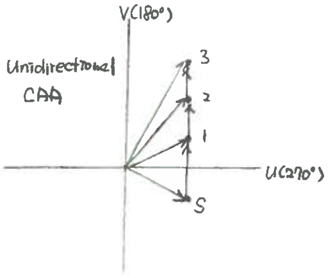
Be sure that the above items are labeled clearly so I can distinguish them. You may draw the vectors directly on the hodograph below

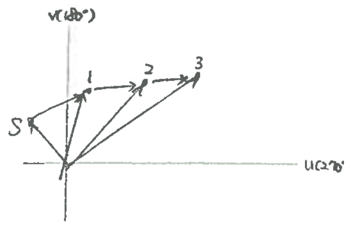
Sol)



3. On your own paper draw two hodographs. One contains <u>unidirectional</u> shear with <u>cold</u> air advection. The second contains <u>clockwise</u> shear. Each hodograph should show the surface wind and winds at altitudes of 1, 2, and 3 km (labeled S, 1, 2, 3). The shears between these levels also should be shown

Sol)





4. List the four physical processes that determine buoyancy at the convective scale.

Sol) ( )* ≡ perturbation variable.

(a) Perturbation temperatures

$T^* > 0$ will lead to positive buoyancy.

(b) Perturbation pressures

$P^* < 0$ positive buoyancy.

(c) Moisture

Moist air is less dense. It will be positively buoyant.

(d) Hydrometeors

The greater the # of hydrometeors ⇒ the less buoyancy

5. a. Explain the process(es) by which horizontal vorticity can be transformed into vertical vorticity.

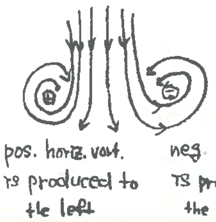
b. A parcel of air is negatively buoyant, e.g. a downdraft. Draw a sketch of streamlines showing the horizontal vorticity that is produced. Make sure that your streamlines have arrows that indicate the sense of the vorticity.

Sol)

(a) Horiz. vort. ⇒ vert. vort.

This may occur due to tilting/twisting of the horizontal vorticity. In addition, horiz. vort may result from baroclinic sources due to horiz. gradients of buoyancy. Then, this horiz. vort. may be transformed into vert. vort. by tilting/twisting

(b)



Horiz. vort. is produced about a horiz. axis running into and out of the page.

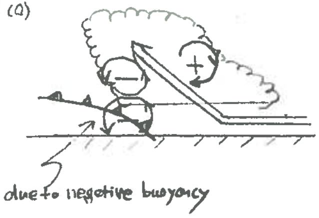
✓ 6. Draw two cross sections that show how a cold pool (gust front) can be influenced by vertical wind shear. Specifically,

a. One sketch shows the gust front with no environmental shear. Use horizontal vorticity concepts to explain whether new cell development is likely

b. The second sketch shows the gust front with low level wind shear. Use horizontal vorticity concepts to explain whether new cell development is likely.

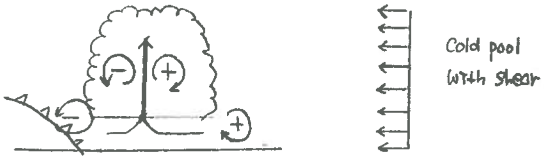
Sol)

(a)



no shear: Cold pool associated with gust front more (−) horiz. vort. Updraft tilts to the left new cell development not likely weaker vertical motions. (air less likely to reach LFC)

(b)



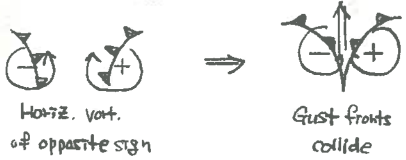
Horiz. vorticities balance. The pos. horiz. vort. due to the shear cancels out neg. horiz vort due to the cold pool. We expect upright convection and maximum vert. motions. New cell development is likely.

7. Draw a cross section through a multicell storm complex. It should be oriented along the direction of storm complex motion. Show the various cells as well as streamlines indicating the flow. Indicate where air is coming into or out of your cross section. Describe the important points of your cross section in several sentences. (Remember that the sketch alone is not sufficient).

Sol) Skip.

8. Use the concept of horizontal vorticity to explain why storms frequently devel when two gust fronts collide — even though the environmental shear is very smal

Sol) Even though environmental shear is very small, colliding gust fronts frequently result in new storm development. This occurs as one of the gust fronts acts as the environmental shear. Vertical motions (upward) are maximized as the gust fronts collide.



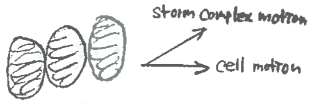
✓ 9. a. Explain the difference between discrete and continuous propagation. Which is dominant in multicell storms? ... in supercell storms?

b. Show a cell motion vector and a .... propagation vector that often are associated with flash flooding. Explain what is going on

Sol)

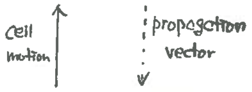
(a) Discrete propagation : new cells form as old cells die

The whole complex moves one direction, while individual cells, which are forming and dying, move in another direction. This is usually found in multicell storm complexes.



Continuous propagation : a single cell will constantly be changing form as one side of the cell is favored over another. This is usually associated with supercells.

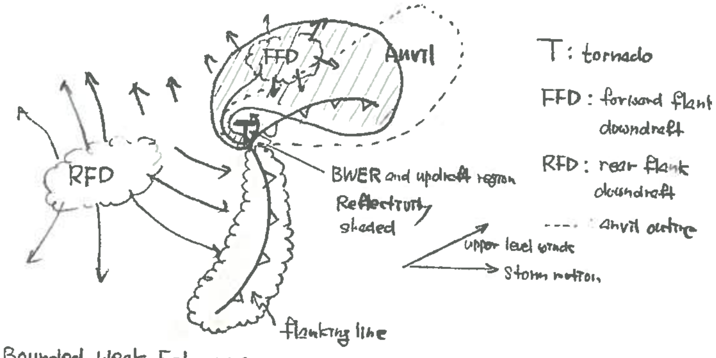
(b)



These two vectors are of equal and opposite strength, and therefore cells may stay over the same location for long periods of time ⇒ training.

✓ 10. Sketch a plan view of a classic tornadic supercell. Indicate the various gust fronts and downdrafts areas, the bounded weak echo region and updraft, the anvil, the flanking line, and the location of the tornado. Label each item carefull

Sol)



Bounded Weak Echo region

✓ 11. One part of the dynamic perturbation pressure eq. (p^*) deals with rotation.

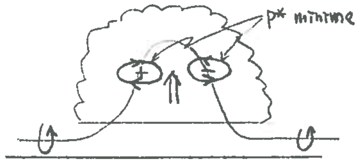
a. Explain this term in a way that you mother understand.

b. What kind of vertical distribution of p^* and resulting p^* force result from this process in a supercell? (You should describe certain aspects of storm structure here.) How does this distribution of p^* affect the evolution of the supercell?

Sol)

(a) The perturbation pressure, p^*, is proportional to the negative of the vert. vort. squared $p^* \propto -\zeta^2$. Therefore, no matter the orientation of the spin (cyclonic or anticyclonic), the p^* will always be negative. A low p^* will develop.

(b)



due to shear horiz. vort. ⇒ tilted ⇒ vert. vort ⇒ p^* minima ⇒ upward directed p^* force ⇒ favored areas for updraft development in time.

✓ 12. There is a second term in the dynamic perturbation pressure eq. that is very important in supercells. It is given by

$$\nabla^2 p_D^* = -2\rho_0 \vec{S}\cdot\nabla w$$

a. Describe this term in your own words

b. Assume you are downshear of a strong updraft. What will be the vertical distribution of p^* and the resulting p^* force in this area? How does this affect the evolution of the updraft?

Sol) Skip.

✓ 13. Describe the sequence of events that leads to the formation of the rear flank downdraft just prior to tornadogenesis. This explanation should be based on vorticity concepts.

Sol)

We have horiz. vort. due to env. shear at first. In addition to this, after the forward flank downdraft has formed, we get additional horiz. vort. due to gradients of buoyancy. We get baroclinically-generated horiz. vort. which is often greater than the horiz. vort. due to the environment shear. Then this vort. gets advected. It is then tilted and streched. We get a downward directed perturbation pressure force usually to the rear of the main updraft. This then leads to the formation of the RFD.

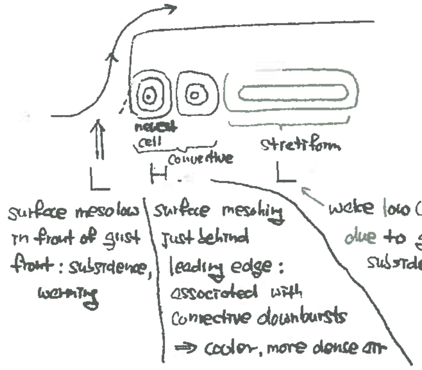
✓ 14. Describe the life cycle of a tornado in words, i.e., the visual aspects of it.

Sol) First, there is an organizing stage. A funnel cloud may form from an organized wall cloud. Debris may soon be seen at the ground before the condensation funnel extends all the way down. The next stage is the mature stage. The tornado strengthens and reaches its maximum width and strength. Then, we reach the shrinking stage. The wall clouds begins to dissipate, while the tornado gets taller. The tornado may tilt due to the interaction with the RFD and FFD. Finally, we reach the decaying stage where there is little or no wall cloud and the tornado may be rope-like

✓ 15. Sketch a cross section through a convective line with a trailing stratiform precipitation area. Indicate the precipitation areas (new/old convective areas and stratiform), the important flow regimes, and the location of meso highs and mesolows at the surface.

Sol)

squall line/MCS moving westward
(tropical or in mid-lats too)



surface mesolow in front of gust front: subsidence, warming

surface meso high just behind leading edge: associated with convective downbursts ⇒ cooler, more dense air

wake low (surface) due to general subsidence; warming.

16. Microbursts have been found to be caused mostly by the buoyancy term in the vertical eq. of motion. Modeling studies have determined the combinations of lapse rate and rainwater that are conducive to burst formation.

a. Describe these findings for wet microbursts and for dry microbursts.

b. Sketch a temperature/dew point sounding for a typical dry microburst case.

Sol)

(a) dry microbursts:

very steep lapse rate needed (close to γ_d)

rainwater < 0.25 mm

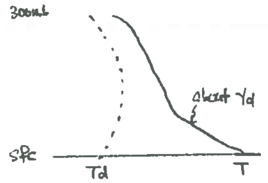
wet microbursts

not as steep of a lapse rate is needed (6 ~ 9°C/km)

rainwater > 0.25 mm

θ_e found to decrease with height more rapidly than normal

(b)



onion-type sounding:

Any parcel cooler than environment will rapidly descend.

Kloesel

- eq. of continuity for Intro. to dynamics
- hydrostatic eq. → hypsometric eq.
- Coriolis force (2 questions)
- Convective storms : Single-cell, Multi-cell, super-cell
- deepening extratropical cyclones
- traveling extratropical cyclones
- Squall line
- Mixed ....

* eq. of continuity

1/

MET? : Kloesel (?,1997)

- On short notice you have been asked to give a lecture in Intro to Dynamics. You need to cover both the mathematical development and the physical interpretation of the terms in equation of continuity. Produce notes for your lecture, including sketches of any diagrams you might show to illustrate relevant aspects of your derivation.

* hydrostatic eq. → hypsometric eq. + application

MET? : Kloesel (20-30 minutes) *

- QUESTION: Scale analysis of the third equation of motion for meso- and synoptic-scale systems results in what is known as the hydrostatic approximation.

A) Use the hydrostatic approximation to derive an expression for the height of the base of any atmospheric layer.

B) Use the expression derived in (A) to determine what effect warming in the lower stratosphere would have on an existing cold-core 500 mb low.

C) Use the expression derived in (A) to determine what effect cooling in the lower stratosphere would have on an existing cold-core 500 mb low.

(List all assumptions you make along the way).

A.

Start with vertical momentum eq. in spherical coord.

$$\frac{\partial w}{\partial t}+\mathbf{V}\cdot\nabla w+w\frac{\partial w}{\partial z}-\left(\frac{u^2+v^2}{a}\right)=-\frac{1}{\rho}\frac{\partial p}{\partial z}-g+2\Omega\cos\phi\, u+F_{rz}$$

$$\begin{array}{lcccccc}
\text{Scale as}\rightarrow & \frac{UW}{L} & \frac{U^2}{a} & \frac{P_0}{\rho H} & g & f_0 U & \nu\frac{W}{H^2} \\
m/s^2\rightarrow & \frac{10(0.01)}{10^6} & \frac{10^2}{6.37\times10^6} & \frac{10^5}{1\cdot10^4} & 10 & 10^{-4}\cdot10 & \frac{10^{-5}\cdot10^{-2}}{(10^4)^2} \\
 & \downarrow & \downarrow & \downarrow & \downarrow & \downarrow & \downarrow \\
 & 10^{-7} & 10^{-5} & 10 & 10 & 10^{-3} & 10^{-15}
\end{array}$$

$$\underbrace{10\quad 10}_{\text{hydrostatic balance}}$$

Thus in arriving at the hydrostatic balance relation we assume that vertical acceleration $\left(\frac{dw}{dt}\right)$, the vertical Coriolis force & frictional dissipation in the vertical are all much less than the vertical pressure gradient and gravitational accelerations. Thus the pressure at any point equals the weight of a unit cross-section column of air above that point.

Now integrate the hydrostatic balance from pressure level P_0 to P_1 (Z_0 to Z_1)

$$\frac{dp}{dz}=-\rho g$$

$Z_0 < Z_1$

$P_0 > P_1$

From eq of state $\rho=\frac{P}{RT}$

so $-\frac{RT}{g}d\ln p = dz$

* We see layer thickness & layer mean temperature

and

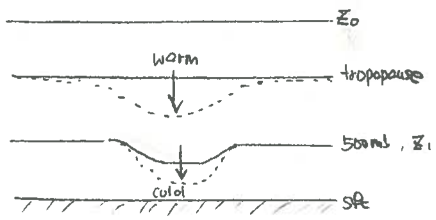
$$-\int_{P_0}^{P_1}\frac{RT}{g}d\ln p=\int_{Z_0}^{Z_1}dz$$

$$\boxed{\frac{R\overline{T}}{g}\ln\left(\frac{P_0}{P_1}\right)=\underbrace{Z_1-Z_0}_{\text{layer thickness}}}\quad \text{where } \overline{T}=\frac{\int_{P_1}^{P_0}T\,d\ln p}{\int_{P_1}^{P_0}d\ln p}\leftarrow \text{layer mean temp}$$

→ hypsometric eq

B.

Schematically we have



Warming in lower stratosphere will decrease static stability by spreading/pushing surfaces apart. The tropopause will dip down. The 500mb low will deepen.

Due to high static stability in stratosphere it will resist vertical displacement of pressure sfc's above the heating. Rather the lower stratosphere layer will tend to expand into the troposphere. Reverse our perspect

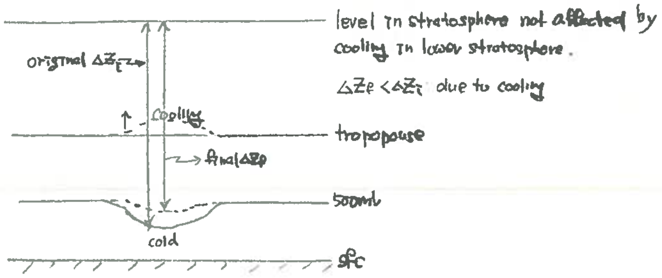
Let Z_0 be some level in stratosphere above the heating whose height does not change due to the heating. Let Z_1 be the height of 500mb sfc. With warming in the lower stratosphere $\overline{T}\uparrow$ $\therefore \Delta Z\uparrow$. Assuming Z_0 remains roughly constant we must see $Z_1\downarrow$ in order that $\Delta Z\uparrow$.

Hence the 500mb cold core low deepens in respond to lower stratospheric warming

C. Lower stratospheric cooling → 500mb low weakens (opposite effect)

$\overline{T}$ decrease $\therefore \Delta Z$ must decrease

Assuming Z_0 does not change due to large static stability in stratosphere above cooling, we must have $Z_1\uparrow$ so the 500mb low weakens.



* "Lower stratosphere warming does not necessarily have to deepen the lower tropospheric low because the warming in the stratosphere could be exactly compensated for by cooling in the lower troposphere"

In this case $\overline{T}_{initial}=\overline{T}_{final}$ and $\Delta Z_{initial}=\Delta Z_{final}$

That is, there is no change in the 500mb height.

If we assume warming/cooling in lower stratosphere is not compensated for by cooling/warming elsewhere in the column, then we should see an effect at 500 mb.

* Note for hypsometric eq. + its interpretation.

Define the geopotential height Z as

$$Z \equiv \frac{\Phi(z)}{g_0} = \frac{1}{g_0}\int_0^z g\,dz \qquad (1)$$

where g_0 is the globally averaged acceleration due to gravity at the earth's surface. We take $g_0 = 9.8$ m/s. The geopotential $\phi(z)$ represents the work that must be done against the earth's gravitational field in order to raise a mass of 1 kg from sea level to the geometric height z above sea level ($z=0$). By convention we let $\Phi(0)=0$.

Assume hydrostatic balance in an atmosphere in which the air is a mix of dry air (with constituent components similar to those found in a volume of dry air on the earth) plus water vapor. Let the specific humidity, q, denote the mass of water vapor to the total mass of the dry air - water vapor mix per unit volume. To account for the effects of water vapor we replace the dry parcel temperature, T, with the virtual temperature, T_v, where

$$T_v \triangleq (1+0.61q)T \qquad (2)$$

The assumed hydrostatic balance is expressed by

$$\frac{dp}{dz} = -\frac{pg}{R_d T_v} \qquad (3)$$

$\uparrow$ use eq. of state $p = \rho R_d T_v$ to replace ρ

Rewrite (3) as

$$g\,dz = -R_d T_v\, d\ln p$$

But by definition $d\phi = g\,dz$ so the above may be rewritten as

$$d\Phi = -R_d T_v\, d\ln p \qquad (4)$$

Now integrate (4) between pressure levels p_1 and p_2, with geopotential ϕ_1 and Φ_2 ($\Phi_2 > \Phi_1$ or equally $P_2 < P_1$). We obtain

$$\Phi_2 - \Phi_1 = -R_d\int_{p_1}^{p_2} T_v\, d\ln p = +R_d \overline{T_v}\ln\left(\frac{P_1}{P_2}\right) \qquad (5)$$

where $\overline{T_v} = \int_{P_2}^{P_1} T_v\, d\ln p \Big/ \ln\left(\frac{P_1}{P_2}\right)$

is the layer mean temperature → the layer between pressures P_1 + P_2.

Divide both sides of (5) by g_0 and make use of (1). We obtain the hypsometric eq.

$$Z_2 - Z_1 = \frac{R_d}{g_0}\overline{T_v}\ln\left(\frac{P_1}{P_2}\right)$$

Consider the thickness of the 1000-500 mb layer.

Here $p_1 = 1000$ mb, $p_2 = 500$ mb. R_d is a constant 287 J/kg K

$g_0 = 9.8$ m/s²

Suppose $Z_2 - Z_1 = Z_{500mb} - Z_{1000mb} = 5400$ m

Then by (6) the corresponding layer mean virtual temperature is

$$\overline{T_v} = (Z_2 - Z_1)\left[\frac{R_d}{g_0}\ln\left(\frac{P_1}{P_2}\right)\right]^{-1}$$

$$= (5400\text{ m})\left[\frac{287\text{ J/K kg}}{9.8\text{ m/s}^2}\ln\left(\frac{1000\text{ mb}}{500\text{ mb}}\right)\right]^{-1}$$

$$\simeq 266.0\text{K}$$

units

$$m\left[\frac{\frac{kg\cdot m}{s^2}\, m\, \frac{1}{K\,kg}}{m/s^2}\right]^{-1}$$

$$\hookrightarrow m\left(\frac{m}{K}\right)^{-1} = K$$

Now suppose $Z_{500mb} - Z_{1000mb} = 5430$ m ← an increase of 30m

$$\overline{T_v} \simeq 267.5\text{K}$$

→ an increase in geopotential height of 30m in the 1000-500mb thickness corresponds to an increase in $\overline{T_v}$ of about 1.5K

Repeat the exercise but with $P_1 = 1000$mb $Z_1 = 0$m

$P_2 = 925$mb $Z_2 = 652.8$m

$$\overline{T_v} \simeq 285.9\text{ K}$$

Now let $Z_2 = 680.8$m $\longrightarrow \overline{T_v} \simeq 298.2$K

For the 1000-925mb layer an increase in the layer thickness of 27.2m (~30m) corresponds to an increase in temperature of ~ 12K.

* Why is the temperature increase for the 1000-925 mb layer much greater (12K vs 1.5 for a similar increase in geopotential height of 30m?

Possible explanations

① Mathematical: The natural log, ln, tends to zero as the argument tends on one (from values greater than one). That is,

$$\lim_{x\to 1^+}\ln(x) = 0$$

Therefore, $\lim_{x\to 1^+}[\ln(x)]^{-1} = \infty$. We have $\overline{T_v}$ as dery ✓ivry proportional to $\ln(P_1/P_2)$ w $P_1 > P_2$. As the ratio $P_1/P_2 \longrightarrow 1$ we expect $\overline{T_v}$ to become extremely sensitive to small changes in the geopotential thickness $Z_2 - Z_1$ due to the inverse $\ln(x)$ relatio between $\overline{T_v}$ and the ratio of the pressure levels. Consider the standard pressure levels

	ratio	ln(ratio)	[ln(ratio)]$^{-1}$
1000 mb			
	$\frac{1000}{925} = 1.08$	0.07696	12.994
925 mb			
	1.09	0.08618	11.604
850			
	1.21	0.19062	5.246
700			
	1.4	0.33647	2.972
500			
	1.67	0.51282	1.950
300			
	1.5	0.40547	2.466
200			
	1.33	0.28518	3.507
150			
	1.5	0.40547	2.466
100			

↑ very sensitive to slight changes in the ratio. But then this sensitivity is characteristic of the log function (and its inverse → the exponential function)

Thus, the same change in geopotential thickness between the 300+500mb layer and the 925+1000mb layer would result in a temp. change in the 925-1000mb layer exceeding that in the 300-500mb layer by about $\frac{12.994}{1.950} \sim 6.664$ times!

② Physical:

In a hydrostatic atmosphere the pressure at any point in a column of unit cross sectional area is directly proportional to the weight of the air in the column above the point in question. For example, there is a greater mass of air above the 925mb surface than the 500 mb surface. We know that pressure (and density) roughly decrease exponentially with height. We also know that the thickness of a layer is directly proportional to the layer mean temperature. If the layer thickness increases, the layer mean temperature increases. If we take the lower level of the layer to be at the sfc an increase in layer thickness implies that the upper level of the surface must move up. The temperature increase reflects an increase in the layer mean internal energy which is necessary to move the upper level upward. Since there is less mass to displace upward as pressure decreases it seems reasonable that the necessary temperature increase also decreases. Since pressure decreases exponentially with increasing height we expect a similarly rapid decrease in the required temperature increase.

* Coriolis force (Merry-go-round)

MET? : Kloesel (?,1997)

- Recently, I heard an Intro. To Meteorology instructor attempting to explain the Coriolis force by appealing to the experience of trying to walk from the center to the edge of a moving merry-go-round. The instructor told the class "This is very difficult to do because of the Coriolis force!" Discuss this explanation.

Sol)

- The Coriolis force is a pseudo force, not a real force. It appears to us in our rotating as a deflective force. The lecturer's explanation is incorrect. The difficulty in walking radially outward on a merry-go-round arises from the conservation of angular momentum. If the person is initially at rest at the center of a platform moving with angular velocity Ω.

⇒ The angular momentum of the person is

$$AM_i = \Omega \underset{r_0}{r^2} = 0$$

Now as the person walks radially outward the distance of the person from the center of the platform to his/her location increases we have $\Omega r^2 \neq 0$. To conserve AM the person's motion must develop a tangential component such that

$$0 = \Omega r^2 + \frac{V_t}{r} r^2$$

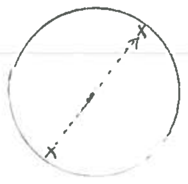
or

$$V_t = -\frac{\Omega r^2}{r} = -\Omega r$$

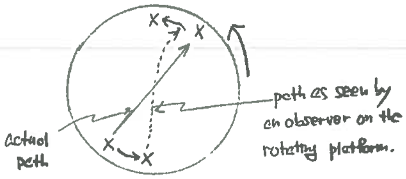
The force the person feels is real and it is due to the conservation of AM. The Coriolis force is not a real force in the sense that it can do no work ⇒ it can change the direction of motion but it can not change the speed of the motion.

- The Coriolis force describes an apparent force that is due to the rotation of the earth. To understand how it works, consider two people playing catch as they sit opposite one another on the rim of a merry-go-round.

ⓐ nonrotating platform



ⓑ rotating platform



If the platform is not rotating, each time the ball is thrown, each person on the platform sees it travel along a straight line.

We now start the platform in motion, say CCW which is the same sense of rotation of the earth in the northern hemisphere when viewed from the North pole. If we watch the game of catch from the stationary ground we see the ball travel in a straight line just as before. However, to the people on the merry-go-round the ball seems to veer to the right each time it is thrown. This is due to the fact that while the ball moves in a straight line path, the merry-go-round rotates beneath it. By the time the ball reaches the other side the catcher has moved. To anyone on the merry-go-round it appears as if there is some force causing the ball to deflect to the right. This apparent force is called the Coriolis force (effect).

This effect occurs on the rotating earth. Viewed from our rotating, non-inertial frame of reference all free moving objects seem to deflect from a straight line path because the earth rotates beneath them.

In summary, to an observer on the earth, objects moving in any direction are deflected to the $\binom{\text{right}}{\text{left}}$ of their intended path in the $\binom{\text{N.H}}{\text{S.H}}$. The amount of deflection depens on

① rotation rate Ω of the earth

② the latitude ϕ of the object

③ the speed V of the object.

$$\left(\frac{\text{Coriolis}}{\text{mass}} \equiv 20\Omega\phi \cdot V\right)$$

The Coriolis force acts perpendicular to the motion

⇒ ∴ It can do no work, only can change direction, not speed.

1/

* Coriolis force

MET? (General): Kloesel (?) *

- Recently, a lecturer in a MET 1010 course attempted to explain the Coriolis force. He appealed to the experience of being ON a merry-go-round and attempting to walk from its center to the edge. "This process is difficult to do", he said, "because of the Coriolis force!" Discuss his explanation.

See the previous question!

★ convective storms : Single-cell, Multi-cell, Super-cell.

MET? (MESO): Kloesel (?) *

- Observed types of convective storms are categorized in three ways. Single-cell, Multi-cell and Super-cell. Explain the background atmospheric conditions which are conducive to the development of each type of convective storm. Include a discussion of the thermodynamic structure and vertical wind shear in your explanation.

Sol)

- Single cell

Single cell storms normally occur in environments where winds are relatively light and vertical shear is small. These storms usually persist for less than one hour, are unsteady, and are relatively small (5-10km). Threats of excessive rainfall are not great due to the short lifetime of the storms.

Cumulus stage : Single cell storms consist of a single updraft which rises rapidly through the troposphere and produces large amounts of liquid water and ice.

Mature stage : When the rain drops and/or ice particles become too heavy for the updraft to support, they begin to fall.

dissipating stage : As the precip. falls it creates a downdraft that, if there is no vertical shear, destroys the updraft underneath it, and then mixes with some of the unsaturated air it encounters on the way down.

The cooling associated with evaporation of downdraft air accelerates the downdraft. The downdraft spreads out horizontally as a cold surge (gust front) on reaching the sfc. The storm system typically moves with the mean environmental wind over the lowest 5-7 km AGL. Severe WX such as high winds or hail may occur but is short lived. Tornadoes are rare.

- Multicell storm

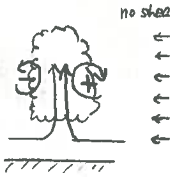
The multicell storm may be viewed as a cluster of short lived single cells. The cold outflows from each cell combine to form a large gust front, the convergence along its leading edge being generally strongest in the direction of storm motion. This convergence triggers new updraft development along and just behind the gust front. The new cell growth often appears disorganized but occasionally occurs on a preferred storm flank where each individual cell moves roughly with the mean wind. The storm's motion as a whole, however may deviate substantially from the mean wind direction due to the discrete redevelopment of cells.

Multicell storms occur in environments of significant buoyant energy (LI ~ -8°C) Low to moderate vertical wind shear is present (~2.5 m/s/km) The wind shear normally shows little CW turning with height but some CCW turning has been noted occasionally.

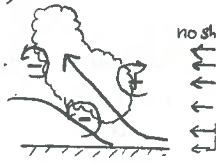
Multicellular storms, especially those that become quasi-stationary, are most likely to develop intense, long-lived convective rainstorms. Thus there exists the possibility of locally heavy rains + flooding. Exceptionally strong updrafts may produce hail. Short lived tornadoes are possible along the gust front in the vicinity of strong updraft centers.

The magnitude of low level (lowest 2km AGL) vertical wind shear affects the ability of a gust front to trigger new cells. Consider 4 cases.

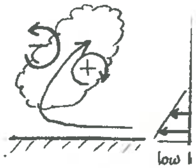
(a) no vertical wind shear and no cold pool ⇒ erect updraft with thermally created, symmetric vorticity



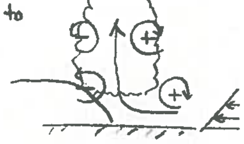
(b) With cold pool and no shear ⇒ distribution is biased by the negative vorticity of the underlying cold pool and cause the updraft to lean in the upshear direction. The ╲ tilt of the updraft means parcels do not reach their LFC near the leading edge of the gust front but rather back among older cells ⇒ not conducive for development.



(c) With low level shear but no cold pool ⇒ updraft tilts toward positive vorticity → a downshear tilt. Precip. falls into low level inflow ⇒ not conducive to storm development.



(d) With low level shear + cold pool ⇒ together combine to yield nearly vertical updrafts towards the front of the gust front. Substantial development.



- Supercell storms

Supercell storms are potentially the most dangerous of convective storm types. It may produce large hail, high winds, and long lived tornadoes over a wide path. In its purest form it consists of a single, quasi-steady, rotating updraft, which may have a lifetime of several hours while propagating continuously to the right (occasionally left) of the mean winds. It is often seen to evolve from multicell storm systems, and even during its quasi-steady phase may comprise several rain centers which evolve similarly to organized multicells. However, the general structure + evolution of the supercell suggests that it is dynamically different from ordinary convection.

Supercell storms usually occur in an environment with strong vertical shear, whe the wind shear vector turns CW (veers) with height below 500mb. Newly formin updrafts continuously feed and reinforce the main updrafts and condense significant amounts of water vapor, the strong winds + wind shear reduce precip. efficiency.

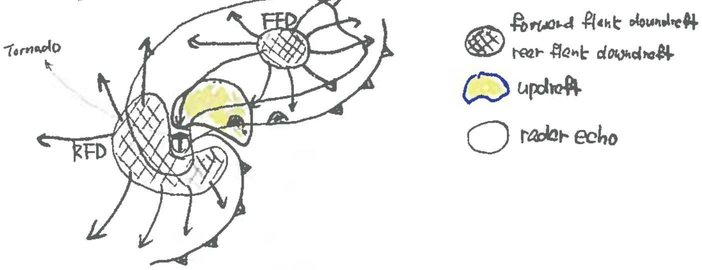
→ Supercell lifecycle

Initial storm development is essentially identical to what would be observed for a short lived single cell or individual cell in a multicell storm complex. The radar reflectivity pattern is vertically aligned and the storm motion is generally in the direction of the mean wind.

1 hour into the storm's lifetime, the reflectivity pattern has elongated in the direction of the mean vertical wind shear vector. The strongest reflectivity gradient is located on the SW flank of the storm. The mid-level reflectivity field also overhangs the low level reflectivity field on the flank of the storm by several km, indicating the presence of a strong updraft. The storm has

now veered to the right of the mean wind.

A supercell usually reaches its mature quasi-steady phase within 90 minutes. A hook like appendage appears to the SW flank of the storm, a large mid-level reflectivity continues to overhang the low level echo. Often a bounded weak echo region appears at mid-levels above the edge of the low-level reflectivity gradient. A BWER usually indicates the presence of both strong updraft and rotation about a vertical axis in its vicinity. Maximum hail fall + tornado occurrence are highest when the BWER collapses. The mesoscale structure of a mature supercell bears striking resemblance to large, synoptic scale cyclones. The strong circulation associated with the rotating supercell wraps the gust front around the southern flank of the storm. This gust front overtakes the frontal boundary associated with the forward flank downdraft. Tornadoes usually form on the top of the occlusion (on the edge of the hook echo) on the gradient between the updraft + downdraft (but within the updraft). A new mesocyclone and updraft may form at the triple point of the occlusion as the old storm center is cut off from its supply of warm air. Not all supercells through this occlusion process but some repeat the sequence several times, leading to a sequential development of tornadoes.



Thermodynamic instability exerts a fundamental control on convective storm strength since it controls the ability of air parcels to accelerate vertically. Vertical wind shear strongly influences the form that convection might take.

① The degree of thermodynamic instability is often measured by easily calculated indices (LI, Showalter, totals totals). Another measure is the convective available potential energy (CAPE)

rising parcel
environment

$$CAPE = \int_{LFC}^{EL} g \frac{T_v(z) - \overline{T}_v(z)}{\overline{T}_v(z)} dz$$

positive area on sounding

$1000-3000 \frac{J}{kg}$ CAPE for moderate to strong convection

CAPE is the buoyancy force integrated w.r.t. height. It represents the work done on the parcel by the environment as the parcel is accelerated upward. Note that this definition of CAPE does not include the effects of pressure perturbations, water loading, and mixing (entrainment). Neglecting these effects the max updraft speed is

$$W_{max} = \sqrt{2 \cdot CAPE}$$

Accounting for these effects reduce W_{max} by ~50%. CAPE calculations are very sensitive to the sfc dewpoint (or mixing ratio in the lowest 500m). An increase in the mixing ratio of 1 g/kg can increase CAPE by 20%

$$\Delta T = 1K \text{ or } \Delta q = 1 g/kg \Rightarrow \Delta CAPE = 200 \sim 600 J/kg$$

CIN ≡ Convective inhibition – measures the strength of the lid inversion

→ the work required to lift an air parcel whose temperature and water vapor mixing ratio are the means in the lowest 500 m from rest at the sfc to the LFC.

② Another important aspect of the thermodynamic structure is the moisture stratification. Large amounts of moisture are needed in the BL to support updraft growth, but the absence of moisture above the BL (2-4km AGL) often enhances storm severity.

→ evaporation in moist downdraft causes it to accelerate ⇒ downburst.

→ Stronger downdraft can lead to stronger gust front ⇒ increase likelihood of updraft redevelopment along strong convergence zone.

Vertical wind shear has a strong influence on the form of convection. Two physical mechanisms might help explain the organizational capacity of vertical wind shear. The first is related to the ability of a gust front to trigger new convective cells. (see early discussion). The second is related to the ability of an updraft to interact with the environmental vertical shear to produce an enhanced, quasi-steady storm structure

→ moderate CAPE + weak vertical wind shear ⟶ single cell.

→ moderate to large CAPE + moderate vertical wind shear ⟶ multi-cell

⇒ redevelopment of cells along a preferred region of an outflow bound

→ high CAPE + large vertical wind shear ⟶ supercell

⇒ updraft + shear interact such that a quasi-steady rotation on the flank of the updraft is developed. This rotation originates through the tilting of horizontal vorticity inherent in the vertically sheared flow.

Since storm behavior is a function of both shear + updraft strength it has been useful to define a bulk Richardson number.

$$Ri_B = \frac{CAPE}{S^2}$$

where $S^2 = \frac{1}{2}(\overline{U}_{6000} - \overline{U}_{500})^2$

↳ mean wind speed in lowest 500m AGL

↳ mean wind speed in lowest 6 km AGL.

Note that S^2 can be interpretated as a measure of the inflow kinetic energy made available to the storm by the vertical wind shear. We've already noted that the K.E. of the updraft may be estimated as

$$\frac{w^2}{2} = CAPE$$

So

$$Ri_B \equiv \frac{\text{potential strength of updraft and indirectly the potential downdraft strength and boundary layer outflow}}{\text{strength of surface inflow feeding the storm and the ability of the storm to take on rotation.}}$$

Observational + numerical studies suggest the following Ri_B ↔ type of convection relationship.

$Ri_B \geq 30$ to 40 ⇒ ordinary cells – the vertical shear is too weak for the updraft. Precip falls into updraft and kills the storm.

$Ri_B \geq 30$ ⇒ multicells

$10 \leq Ri_B \leq 40$ ⇒ supercells → however, if CAPE + S^2 are both small, long lived cells are possible but not likely to be severe.

It is not advisable to forecast storm type based solely on Ri_B!!!

Now a brief discussion of pressure perturbation effects in supercells.

Bluestein derives the following diagnostic perturbation pressure eq in the Boussines system.

$$\nabla^2 p' = -\overline{\rho}\overset{\textcircled{a}}{\left[\left(\frac{\partial u'}{\partial x}\right)^2 + \left(\frac{\partial v'}{\partial y}\right)^2 + \left(\frac{\partial w'}{\partial z}\right)^2\right]} - 2\overline{\rho}\overset{\textcircled{b}}{\left[\frac{\partial u'}{\partial y}\frac{\partial v'}{\partial x} + \frac{\partial w'}{\partial z}\frac{\partial w}{\partial x} + \frac{\partial v'}{\partial z}\frac{\partial w'}{\partial y}\right]}$$

$$-2\overline{\rho}\overset{\textcircled{c}}{\left[\frac{\partial \overline{U}}{\partial z}\frac{\partial w'}{\partial x} + \frac{\partial \overline{V}}{\partial z}\frac{\partial w'}{\partial y}\right]} + \overline{\rho}\overset{\textcircled{d}}{\frac{\partial B}{\partial z}}$$

where ⓐ is the nonlinear fluid extension term ⟶ $P'_{NL_{EX}}$

ⓑ is the nonlinear shear term ⟶ P'_{NL_S}

ⓒ is the linear term ⟶ P'_L

ⓓ is the buoyancy term ⟶ P'_B

ⓐ+ⓑ+ⓔ ≡ the pressure perturbation associated with gradients in the wind field.

→ refered to as the dynamic pressure

→ ~~decompose into nonlinear extension & shear terms~~ plus a linear term

ⓓ ≡ the pressure perturbation associated with the vertical derivative of buoyancy → the buoyancy pressure.

We rewrite ⓒ as $-2\bar{\rho}\frac{\partial \bar{V}}{\partial z}\cdot\nabla w'$. Then $\nabla^2 P_L' \propto -P_L'$

so

$$P_L' \propto 2\bar{\rho}\frac{\partial \bar{V}}{\partial z}\cdot\nabla w'$$

At a given level there is a negative perturbation pressure on the downshear side of an updraft and a positive perturbation pressure on the upshear side of an updraft.

Strong unidirectional shear from the sfc to mid levels coupled with a strong, localized updraft is important ⇒ through the dynamics represented by P_L' this can promote new cell growth on the forward, downshear side of the updraft and suppress new cell growth on the upshear side.

Given the same speed shear but with the addition of directional shear we get preferred development on the left or right flank of the storm.

(a) CW shear favors upward pressure perturbations on the right flank

(b) CCW shear favors left flank development.

(c) unidirection shear favors downshear development $\left\langle\begin{array}{l}\nearrow \text{can lead to} \\ \searrow \text{splitting cells.}\end{array}\right.$

We see that linear pressure perturbation effects (tilting horizontal vorticity into the vertical) biases the cell movement left, right, or downstream.

Now consider the effect of nonlinear pressure perturbations. Ignore the fluid extension term and make a few more assumptions (see Bluestein, p 468) to arrive at

$$\nabla^2 P_{NL}' = \frac{\bar{\rho}}{2}\underbrace{\zeta'^2}_{\text{(perturbation vorticity)}^2}$$

or

$$P_{NL}' \sim -\zeta'^2$$

Thus the nonlinear shear terms act to produce low perturbation pressure in the vicinity of midlevel cyclonic & anticyclonic vorticity. An upward directed perturbation pressure gradient develops below the level at which the vortices are most intense (usually at midlevels) on each flank of the old updraft, along a line normal to the vertical shear vector (p 469)

↳ The nonlinear shear effect promotes new or continued cell growth on the flanks of (alongside) the old cell.

*deepening extratropical cyclones

MET? (Synoptic): Kloesel (?) *

- In rapidly deepening extratropical cyclones, it is often observed that the lower troposphere cools in the immediate vicinity of the low center. In fact, cooling even occurs in the region of warm advection. Yet, from the hydrostatic approximation, we know that for a low to deepen, the atmosphere must warm.

(i) Explain how the lower troposphere cools in spite of warm advection, and how the pressure still falls when the lower troposphere is cooling?

(ii) How does moisture affect the deepening process.

Sol)

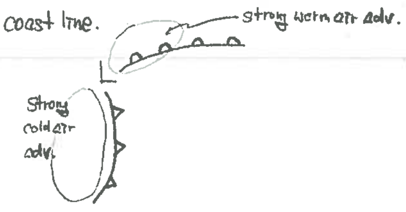
Explosive cyclogenesis occurs most frequently over the ocean during the cold season, downstream from mobile, diffluent, upper-level troughs, within or poleward of the maximum westerly current, and near the strongest SST gradients such as the northern edge of the Gulf Stream + Kuroshio. Formally, "bombs" are cyclones with deepening rates in excess of 1 mb/hr for ≥ 24 hrs (Sanders + Gyakum, 1980).

↳ this pressure tendency is referred to as a Bergeron.

Observations regarding bombs.

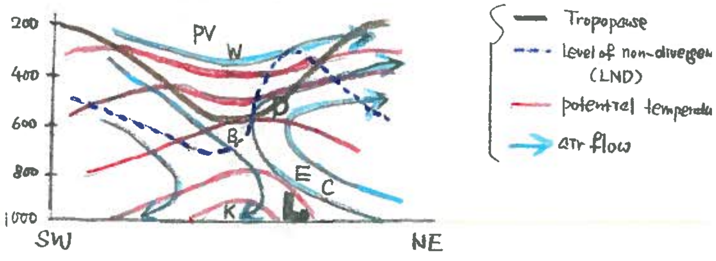
① The turbulent transfer of sensible heat between the relatively warm ocean + cold continental air can create very low static stability near the sfc.

② A temperature gradient is found at the sea surface which can enhance low level temperature advection. Bombs tend to develop and remain along the boundaries of strong sea surface + air temp. gradients suggesting the importance of temperature advection, particularly the warm advection in the lower troposphere. Typically the warm front remains very close to the land-sea boundary where the sfc cyclone creates the most intense pressure falls due to warm advection on the north side of the storm. Cold advection occurs on the equatorward south side of the storm close to the coast line.



③ Boyle + Bosart (1986) found usually large values of temperature adv. at high levels as a sfc cyclone explosively deepened. Quasi-horizontal temp adv. is usually strongest near the ground and weakest aloft. The unusually intense cold air adv. and warm air adv. upstream and downstream, respectively, from an upper level trough were due to the steep slope of the tropopause. The tropopause had been advected unusually far downward by strong subsidence upstream from the upper level trough and west of the sfc cyclone. The tropopause had been advected down to 600 mb, as evidenced by high values of Ertel's potential vorticity, subsidence, and dry air. Thus, relatively warm, stable air is located aloft, while relatively cool air is located near the sfc in the vicinity of the trough axis. The flow through the upper level trough in the presence of the stratospheric air in the region of the low tropopause results in a strong temp adv. pattern. Cyclonic vorticity is generated aloft through convergence above the region of maximum sinking motion. This vorticity is subsequently advected downstream and the differential vorticity forcing is enhanced. Warm air advection downstream from the upper level trough is superimposed over the warm advection below, so that there is a deep column of rising air. The result is a deep layer of rising motion, convergence, and increase in vorticity. The column of convergence aloft associated with the region of subsidence is not so deep because the tropopause, which acts as a lid to vertical motion, is very low. It therefore appears as if the region of subsidence upstream from an upper level trough can, in some instances, enhance surface cyclogenesis.

These features are illustrated in the following diagram



dynamically coupled through "PV thinking"

PV – Stratospheric PV resevior
W – upper-tropospheric warm air
K – lower tropospheric cold air
L – surface Low

- ascending air has a component into the paper.
- descending air has a component out of the paper.

B – production of cyclonic vorticity due to vortex tube stretching accompanying subsidence.

C – region of max lower tropospheric warm air advection

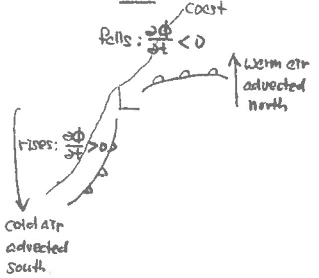
D – vorticity generation due to horizontal convergence beneath elevated LND + surface low.

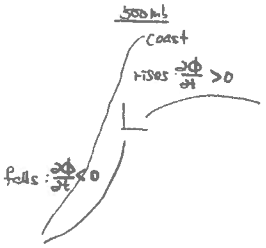
E – parcels subsiding from B gain additional vorticity by convergence as they are advected downstream across the steeply inclined LND.

- Surface cyclogenesis commences as PV max aloft begins to cross the boundary of the cold air dome K while moving toward warmer air.
- Steeply inclined tropopause over L results in significant warm air advection in 400-200 mb layer ⇒ Contributes further to the intensity of the development
- during intense phase the tropopause + PV max lower to the mid troposphere (~600
- high LND over L ensures convergence + cyclonic vorticity generation through a deep layer accompanying the development of a cutoff vortex.

The synoptic pattern in which these storms form is equatorward of a large, cold high at the sfc situated north of the incipient cyclone center. This configuration allows easterly winds north of the storm center to move along an extended fetch from the warmest ocean temperatures towards the coast. Surface fluxes of sensible heat and moisture serve to maintain the high θ_w of the air which feeds into the warm sector of the storm. The high θ_w air enables a low static stability to develop over the lower troposphere. Enhanced condensation is possible when the high θ_w air ascends into the storm ⇒ thereby producing a relatively unstable lapse rate over the lower troposphere.

Movement and development of rapidly developing coastal storms are strongly affected by the low level ∇T, the moisture supply in the form of high θ_w, and the static stability. The 500mb ζ pattern is modified during coastal development such that the vort. max tends to follow the coastline. Since the coastal cyclone is advecting $\begin{pmatrix}\text{warm}\\ \text{cold}\end{pmatrix}$ air $\begin{pmatrix}\text{north}\\ \text{south}\end{pmatrix}$ of the center, geopotential height falls at the sfc and rises at 500mb tend to be concentrated north of the storm along the coastline. Geopotential height rises at the sfc and falls at 500mb tend to be concentrated along the coast south of the sfc cyclone.





* traveling extratropical cyclones (baroclinic vort. eq.)

MET? (Synoptic): Kloesel (?) *

- Use the baroclinic vorticity equation, and your knowledge of the structure of "short waves", to argue why low-level convergence and upper-level divergence develop ahead of traveling extratropical cyclones.

Sol)

Weakness of equivalent barotropic model is that does not include the dynamics necessary to model the development or decay of weather systems. The defficiency is linked to the neglect of thermal (thickness) advection in the model.

Inclusion of thermal advection leads to a two parameter model ⇒ baroclinic model

we assume

① orientation of isotherms is constant with height

② 〃 geostrophic wind shear is constant with height

Define a thermal wind $\underline{V}_T$ + thermal vorticity ζ_T by

$$\underline{V}_g = \underline{V}_m + B(p)\underline{V}_T \qquad \underline{V}_T = \underline{V}_g - \underline{V}_m$$

$$\zeta_g = \zeta_m + B(p)\zeta_T \qquad \zeta_T = \zeta_m - \zeta_0$$

↑ vertical structure function ↑ sfc

└ values at mean level where absolute vorticity is approx. conserved (500 to 600 mb)

Integrate the QG vort. eq. from [P_s, 0] and normalize by P_s to get eq. valid at the mean level (m)

$$\frac{1}{P_s}\int_0^{P_s}\left[\frac{\partial\zeta_g}{\partial t} + \underline{V}_g\cdot\nabla_p(\zeta_g+f)\right]dp = \frac{f_0}{P_s}\int_0^{P_s}\frac{\partial\omega}{\partial p}dp \quad \begin{pmatrix}\omega(0)=0 \text{ as B.C.}\\ \omega(P_s)=\omega_s\end{pmatrix}$$

Substitute for $\underline{V}_g$ and ζ_g. Integrate noting that

① $B_m = \frac{1}{P_s}\int_0^{P_s} B(p)dp = 0$

② ζ_T + $\underline{V}_T$ are const. with pressure

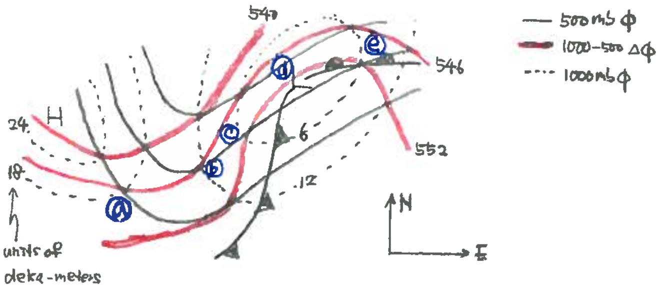
③ ζ_m + $\underline{V}_m$ are const. with pressure

to get

$$\boxed{\frac{\partial\zeta_m}{\partial t} = -\underline{V}_m\cdot\nabla_p(\zeta_m+f) - \overline{B(p)^2}\,\underline{V}_T\cdot\nabla_p\zeta_T + \frac{f_0\,\omega_s}{P_s}}$$

↳ this is the baroclinic vorticity eq.

Consider a typical wave/cyclone pattern shown below.



ⓐ max ζ_T

ⓑ max ζ_g at 500 mb

ⓒ max cold air thickness advection by 1000 mb $\underline{V}_g$

ⓓ minimal thermal vorticity ζ_T lies between ⓓ and ⓔ

ⓔ max warm thickness advection by 1000 mb $\underline{V}_g$.

Max cold air advection occurs near ⓒ. Warm thickness advection by $\underline{V}_g$ (1000 mb) correlates with negative thermal vorticity advection NE of the sfc L_0 near ⓔ and developmental height rises at 500 mb. Max cold thickness advection by $\underline{V}_g$ (1000 mb) corresponds closely in location to the max positive geostrophic thermal vorticity advection by the thermal wind and to 500m developmental height falls.

In the case of a sfc cyclone lying ahead of a 500 mb trough, the regions between the sfc trough and the 500 mb trough, and between the sfc trough and the downstrea 500 mb ridge, experience temperature advection leading to developmental height changes at both the sfc and at 500 mb. Negative $\frac{\partial\Phi}{\partial t}\Big|_{500\,mb}$ and $\frac{\partial P_s}{\partial t} > 0$ occur west + SW of the sfc L_0 in association with cold air advection. Positive $\frac{\partial\Phi}{\partial t}\Big|_{50}$ and $\frac{\partial\zeta_g}{\partial t}\Big|_{500} < 0$ occur NE of the sfc L_0 in association with warm advection.

These principles are illustrated in the 3D schematic below

+ : rising, − : falling } sign of height/pressure tendency.

NVA ≡ negative vort. adv.

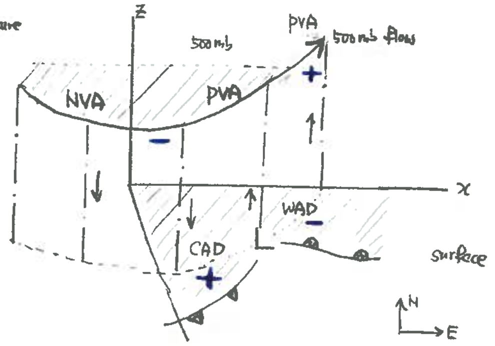
PVA ≡ positive vort. adv.

CAD ≡ cold (air thickness) adv.

WAD ≡ warm (air thickness) adv.

↑ : rising motion

↓ : sinking motion



Max developmental 500 mb Φ height falls occur just ahead of the 500 mb trough and to the rear of the sfc cold front where there is large scale descent + sfc pressure rises. Near ⓒ (see privious diagram) absolute vorticity advection at 500 mb and cold air advection are juxtaposed so that there is both rising motion and cold air advection. Sfc pressure falls, 500 mb Φ rises, and PVA at 500 mb characterize the region east of the sfc cyclone. A similar statement can be made regarding 500 mb Φ rises and absolute vorticity decreases associated with development over the regions east of the sfc cyclone and within the 500 mb ridge.

Sutcliffe was the first to describe development in terms of a vertical distribution of divergence, specifically at the sfc, and readily measurable quantities at 1000 + 500 mb.

Sutcliff first arrived at a simplified form of the QG-ζ eq. He applied this eq at 1000 + 500 mb; subtracted them; and obtained a thickness tendency eq. He then argued + scaled his way to a diagnostic eq for 1000 mb $\nabla\cdot\underline{V}$

$$-\nabla_p\cdot\underline{V}_0 = -\frac{1}{f_0}\left(\underline{V}_T\cdot\nabla(2\zeta_0+f)\right) - \frac{1}{f_0}\underline{V}_T\cdot\nabla\zeta_T$$

In the above

$-\frac{2}{f_0}\underline{V}_T\cdot\nabla_p\zeta_0$ ≡ steering effect. Vertical wind shear (horizontal $\nabla_p T$) over a sfc max of vorticity (manifested by a thickness line crossing the feature) is associated with cyclonic development ahead of the feature + anticyclonic development behind it, resulting in a displacement of the feature in the direction of $\underline{V}_T$.

⇒ cyclones move in the direction of thickness lines over the center of the low (warm air to the right of the motion vector)

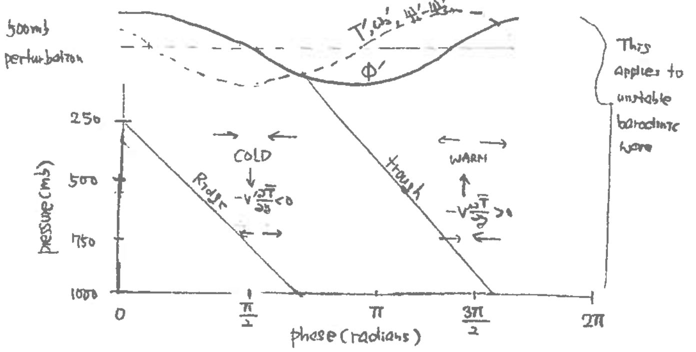
$-\frac{1}{f_0}\mathbf{V}_T\cdot\nabla_p f \equiv \beta$ effect which is generally negligable.

$-\frac{1}{f_0}\mathbf{V}_T\cdot\nabla_p\zeta_T \equiv$ thermal wind advection of thermal vorticity.

This term operates to create low-level convergence and spin up the sfc low if the thermal trough lays behind the sfc low.

Sutcliffe's development eq. is important because it shows that the sfc lows move from one location to another not as bubbles in a stream but rather because the sfc pressure pattern is being continuously reconstituted due to forcing. That is, a sfc low moves because it is continuously being filled up behind + deepened ahead of its center due to the divergence + convergence patterns associated with geostrophic advections at higher levels.

The figure below (Holton, p225) schematically shows the relationship between a geopotential field and the divergent secondary circulation for a developing baroclinic wave for the usual mid-lat situation where $U_T > 0$



At 500mb the thickness field $(\Phi_1'-\Phi_3')$ lags the geopotential field (Φ') by one quarter wavelength. Thickness + vertical motion are in phase. Note that temperature advection by the perturbation meridional wind V' is in phase with the 500mb thickness field so that the advection of the basic state temperature by the perturbation wind acts to intensify the perturbation thickness field. The divergence pattern associated with the subsiding branch of the vertical circulation contributes a positive vorticity tendency at the 250mb trough and a negative ζ tendency at the 750mb ridge. Conversely in the region of ascending motion the divergence field at the 250mb ridge contributes to a negative ζ tendency and a positive tendency at the 750mb trough. Since in all cases these vorticity tendencies tend to increase the extreame values of vorticity at the troughs and ridges, this secondary circulation system will act to increase the strength of the disturbances.

↓

Recall the QG vort. eq. in the form

$$\frac{\partial\zeta_g}{\partial t} = -\mathbf{V}_g\cdot\nabla(\zeta_g+f) - f_0(\nabla\cdot\mathbf{V}_H)$$

We see that neglecting vort. advection

(a) horizontal divergence $(\nabla\cdot\mathbf{V}_H>0)$ forces $\frac{\partial\zeta_g}{\partial t}<0 \longrightarrow \zeta_g\downarrow$

(b) " convergence $(\nabla\cdot\mathbf{V}_H<0)$ " $\frac{\partial\zeta_g}{\partial t}>0 \longrightarrow \zeta_g\uparrow$

From this eq and the hydrostatic, adiabatic thermodynamic eq. for a two-level (baroclinic) atmosphere one can derive an eq. for the vertical velocity at the mean level m. In words, this eq is of the form

(2nd order operator) ω_m = − laplacian of (perturbation thickness adv. by $\mathbf{V}_m$)
+ laplacian of (mean thickness adv. by $\mathbf{V}_T$)
+ differential adv. of perturbation vort. by the basic state wind.

From this eq we can show

$$-\omega_m' \propto \underbrace{-V_m'\frac{\partial\overline{T}}{\partial y}}$$

↳ advection of base state temp. field by the perturbation mean meridional wind.

and

$$-\omega_m' \propto -U_T\frac{\partial\zeta_m'}{\partial x}$$

↳ advection of perturbation mean vorticity by the basic state thermal wind.

We summerize as follows

(a) $\begin{pmatrix}\text{warm}\\ \text{cold}\end{pmatrix}$ temp. adv. forces $\begin{pmatrix}\text{rising}\\ \text{sinking}\end{pmatrix}$ motion

(b) $\begin{pmatrix}\text{negative}\\ \text{positive}\end{pmatrix}$ vort. adv. forces $\begin{pmatrix}\text{rising}\\ \text{sinking}\end{pmatrix}$ motion.

* squall line

MET? : Kloesel (?,1997)

- Doppler radar observations of squall line structure have led to a conceptual model of a squall line as proposed by Houze et al., (1989). Draw this conceptual model in a vertical cross section and discuss the mechanisms for:

a) the trailing stratiform region
b) descending rear inflow
c) ascending rear outflow
d) the 'bright band' as depicted on radar imagery

* Mixed

* Dewpoint temp. + condensation + advection (QG)

MET? : Kloesel (?)

- Answer EITHER 1a OR 1b, then answer 2

1. a. While playing the game "Trivial Pursuit", a card was drawn which contained the following question..

   "At what temperature does the condensation of water vapor occur?"
   ANSWER: DEWPOINT

   Discuss at least <u>one</u> other plausible answer for this not so "trivial" question!

   b. Describe the ways in which the Earth's oceans affect synoptic scale weather systems (in terms of development, maintenance, and decay).

2. Consider the advective rate of change of a scalar quantity '*M*' by the geostrophic wind: $\mathrm{Adv}_M = -\mathbf{V}_g \bullet \nabla M$
   a) Show that Adv_M may be expressed in terms of the Jacobian of the geopotential and '*M*'.
   b) Sketch a pattern of isolines of geopotential and '*M*' that shows both a maximum of positive Adv_M and a minimum of negative Adv_M and indicate the locations of these. Also indicate where $\nabla^2 \mathrm{Adv}_M$ is maximum positive and minimum negative.
   c) Briefly discuss why the above are important in relation to synoptic scale weather systems.

1.a

The question does not specify how we bring the moist parcel of air to saturation (and hence, condense water vapor)

w.r.t. saturation (water vapor saturation); 4 saturation temps are

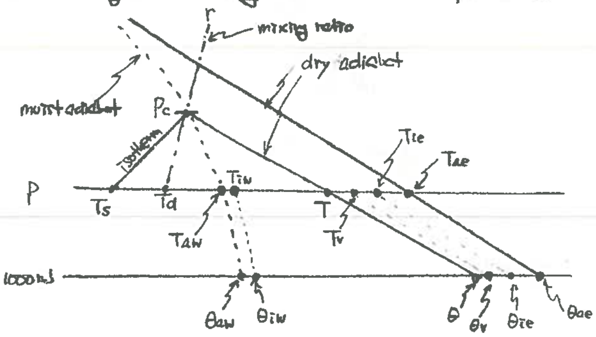
① $T_d \equiv$ dewpoint temp. → temp. at which saturation occurs when air is cooled isobarically.

② $T_{iw} = T_w \equiv$ (isobaric) wet bulb temperature → the temperature which air attains when water is evaporated into it until saturation is reached. This process is done such that pressure remains constant (isobaric) + the air-water system does not exchange heat with the environment (adiabatic)

③ $T_{aw} \equiv$ adiabatic wet bulb or pseudo-adiabatic wet bulb temp. → adiabatically lift a parcel to saturation. moist adiabatically lower the parcel to its original pressure level. In this lowering process we add water vapor to the parcel to maintain a saturated parcel. $\underline{T_{aw} < T_{iw}}$ because to arrive at T_{aw} we have water-air at varying temperatures T_s to T_{aw}. The evaporation of water at each stage in the process in descent to T_{aw} subtracts some heat from the air. T_{iw} was arrived at from water already at T_{iw}.

④ $T_s \equiv$ saturation temp → the temperature a parcel of air attains if it is lifted dry adiabatically to saturation

Below is a diagram illustrating these and other temperatures.



* Normand's Rule

$$\underbrace{T_d \leq T_w \leq T}$$ → equality when air is saturated.

1.b

Land-ocean contrasts serve as the basis for many planetary + synoptic scale motions. The difference in the specific heat and albedo of the land surface and ocean surface lead to differential heating (a covariance between heating and temp.; At the planetary scale this can enhance the baroclinicity of the atmosphere. Differential heating drives many planetary circulations (monsoon circulations). It also leads to mesoscale circulations such as land/sea breezes. Warm oceans (>26°C) are the regions in which hurricanes develop. Hurricanes can transform into extratropical systems as they move into the mid-lats. Hurricanes weaken as they move over land due to increased sfc friction and the loss of their moistu supply. Midlatitude systems can experience explosive development when they move over warm waters such as the Gulf Stream. This type of development is most pronounced during the winter season.

EX :• upper level PVA ahead of 500 mb trough leads to height falls & rising motion. The secondary circulations (vertical motion, low level convergence) develop to maintain the atmosphere in geostrophic & thermal wind balance.

• lower level warm air adv. (localized in a frontal-type zone) leads to rising moti

Do note that advection alone is not the reason systems amplify. Advection transpo quantities around. Certain configuration lend themselves towards development & decay

2.

(a)

The geostrophic wind $\underline{V}_g = \underbrace{-\frac{1}{f}\frac{\partial\Phi}{\partial y}}_{\hookrightarrow u_g}\,\underline{i} + \underbrace{\frac{1}{f}\frac{\partial\Phi}{\partial x}}_{\hookrightarrow v_g}\,\underline{j}$

The gradient of M, $\nabla M = \frac{\partial M}{\partial x}\underline{i} + \frac{\partial M}{\partial y}\underline{j}$

Thus

$$\text{adv}(M) = -\left(-\frac{1}{f}\frac{\partial\Phi}{\partial y}\right)\frac{\partial M}{\partial x} - \frac{1}{f}\frac{\partial\Phi}{\partial x}\frac{\partial M}{\partial y}$$

$$= -\frac{1}{f}\left(\frac{\partial\Phi}{\partial x}\frac{\partial M}{\partial y} - \frac{\partial\Phi}{\partial y}\frac{\partial M}{\partial x}\right)$$

But by definition the Jacobian of scalars α & β is given as

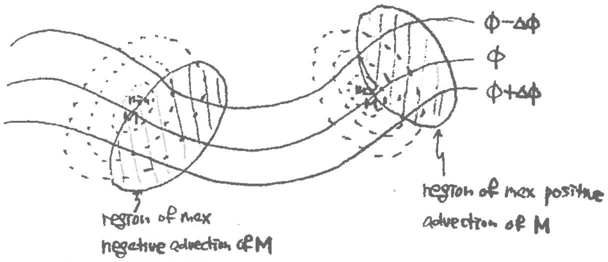
$$J(\alpha,\beta) = \frac{\partial\alpha}{\partial x}\frac{\partial\beta}{\partial y} - \frac{\partial\alpha}{\partial y}\frac{\partial\beta}{\partial x}$$

We identify $\alpha = \Phi$ & $\beta = M$

Then

$$\underline{\text{adv}(M) = -\frac{1}{f}J(\Phi, M)}$$

(b)



The Laplacian operator ∇^2 is such that $\nabla^2\alpha \sim -\alpha$ for scalar α (assuming α is an oscillatory function)

Thus, $\nabla^2 \text{Adv}(M) \sim -\text{Adv}(M)$

Therefore in the above where we have max positive Adv(M), ∇^2Adv(M) is max negative. Similarly max negative Adv(M) is also max positive ∇^2adv(M).

(c)

Importance of the above to synoptic scale weather systems.

Suppose we let $M = \zeta + f$ (absolute vorticity (geostrophic value))

From QG dynamics we derive the geopotential tendency and ω-eqs.

$\chi = \frac{\partial\Phi}{\partial t}$ ≡ geopotential tendency

$$\left[\nabla^2 + \frac{\partial}{\partial p}\left(\frac{f_0^2}{\sigma}\frac{\partial}{\partial p}\right)\right]\chi = -f_0\underline{V}_g\cdot\nabla\left(\frac{1}{f_0}\nabla^2\Phi + f\right) - \frac{\partial}{\partial p}\left[-\frac{f_0^2}{\sigma}\underline{V}_g\cdot\nabla\left(-\frac{\partial\Phi}{\partial p}\right)\right]$$

ω-eq.

$$\left[\nabla^2 + \frac{\partial}{\partial p}\left(\frac{f_0^2}{\sigma}\frac{\partial}{\partial p}\right)\right]\omega = -\frac{f_0}{\sigma}\frac{\partial}{\partial p}\left[-\underline{V}_g\cdot\nabla\left(\frac{1}{f_0}\nabla^2\Phi + f\right)\right] - \frac{1}{\sigma}\nabla^2\left[-\underline{V}_g\cdot\nabla\left(-\frac{\partial\Phi}{\partial p}\right)\right]$$

In words

geopotential $\begin{pmatrix}\text{fall}\\ \text{rise}\end{pmatrix} \propto \begin{pmatrix}+\\ -\end{pmatrix}$ vorticity adv. + $\begin{pmatrix}\text{cold}\\ \text{warm}\end{pmatrix}$ adv. decreasing with height.

$\begin{pmatrix}\text{rising}\\ \text{sinking}\end{pmatrix}$ motion $\propto$ rate of increase with height of $\begin{pmatrix}+\\ -\end{pmatrix}$ vorticity adv.

+ localized $\begin{pmatrix}\text{warm}\\ \text{cold}\end{pmatrix}$ adv.

The advection of scalar quantities (such as vorticity or temp) are important within the framework of QG dynamics in the development, maintenance & decay of midlatitude baroclinic systems