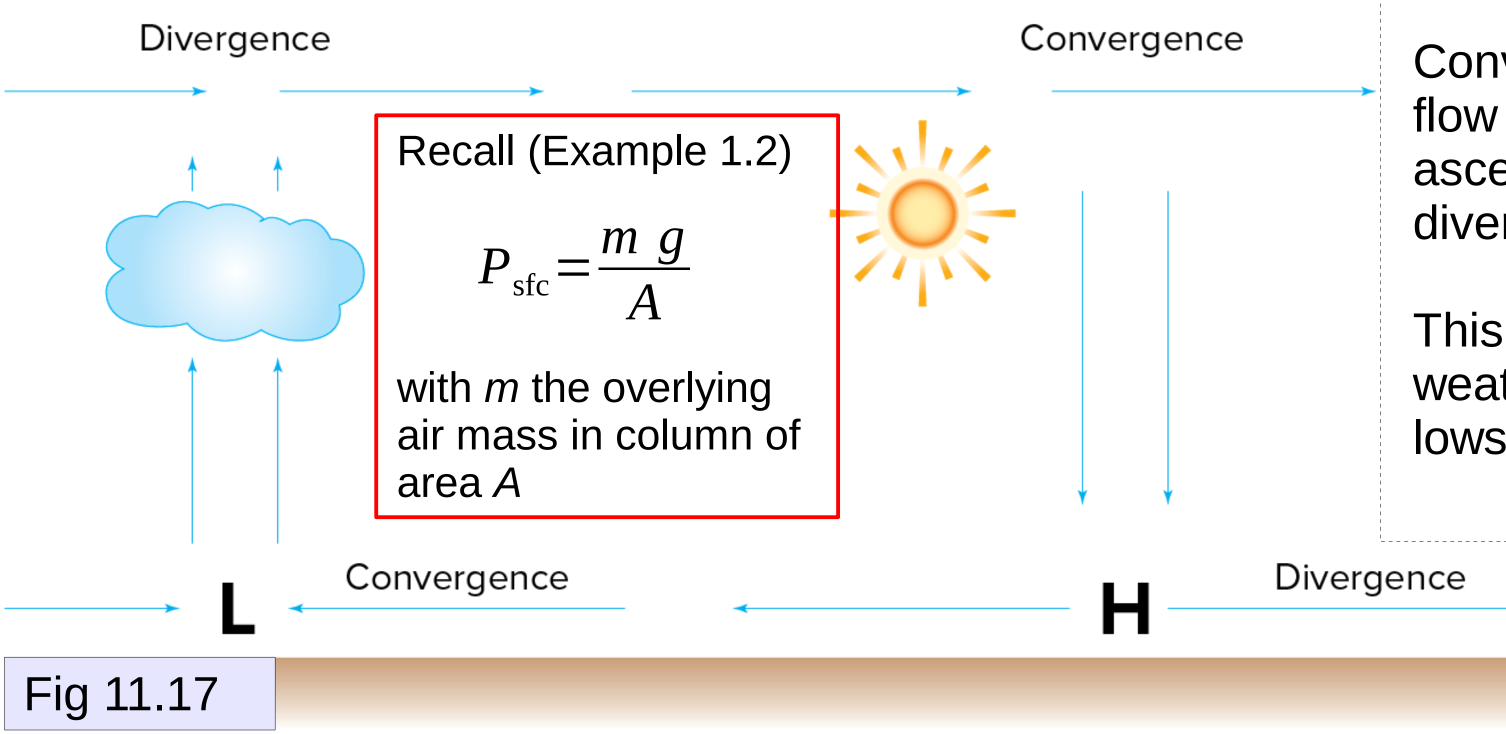
EAS270_Ch11_Winds_B.odp
JDW, EAS U.Alberta, last mod. 21 Nov. 2016

PBL or "friction layer" (Planetary/Atmos. Bound. Layer)

- variation with height above the ABL due to non-constant slope of the isobaric surfaces – the "thermal wind," and its orientation relative to thickness lines
- link between a "veering" or "backing" wind aloft, and thermal advection
- variation with height across the ABL, due to friction

If upper level divergence exceeds surface convergence, the sfc low will deepen



Recall (Example 1.2)

$$P_{\mathrm{sfc}} = \frac{m\, g}{A}$$

with *m* the overlying air mass in column of area *A*

Fig 11.17

Result of coordination of surface pressure system and waves aloft:

Converging cross-isobar flow at the surface feeds an ascending current with divergence aloft

This explains why bad weather is associated with lows

Let x be normal to contours of Z

$$V_g = \frac{g}{f_c} \left| \frac{\Delta Z}{\Delta x} \right|$$ is the geostrophic wind speed

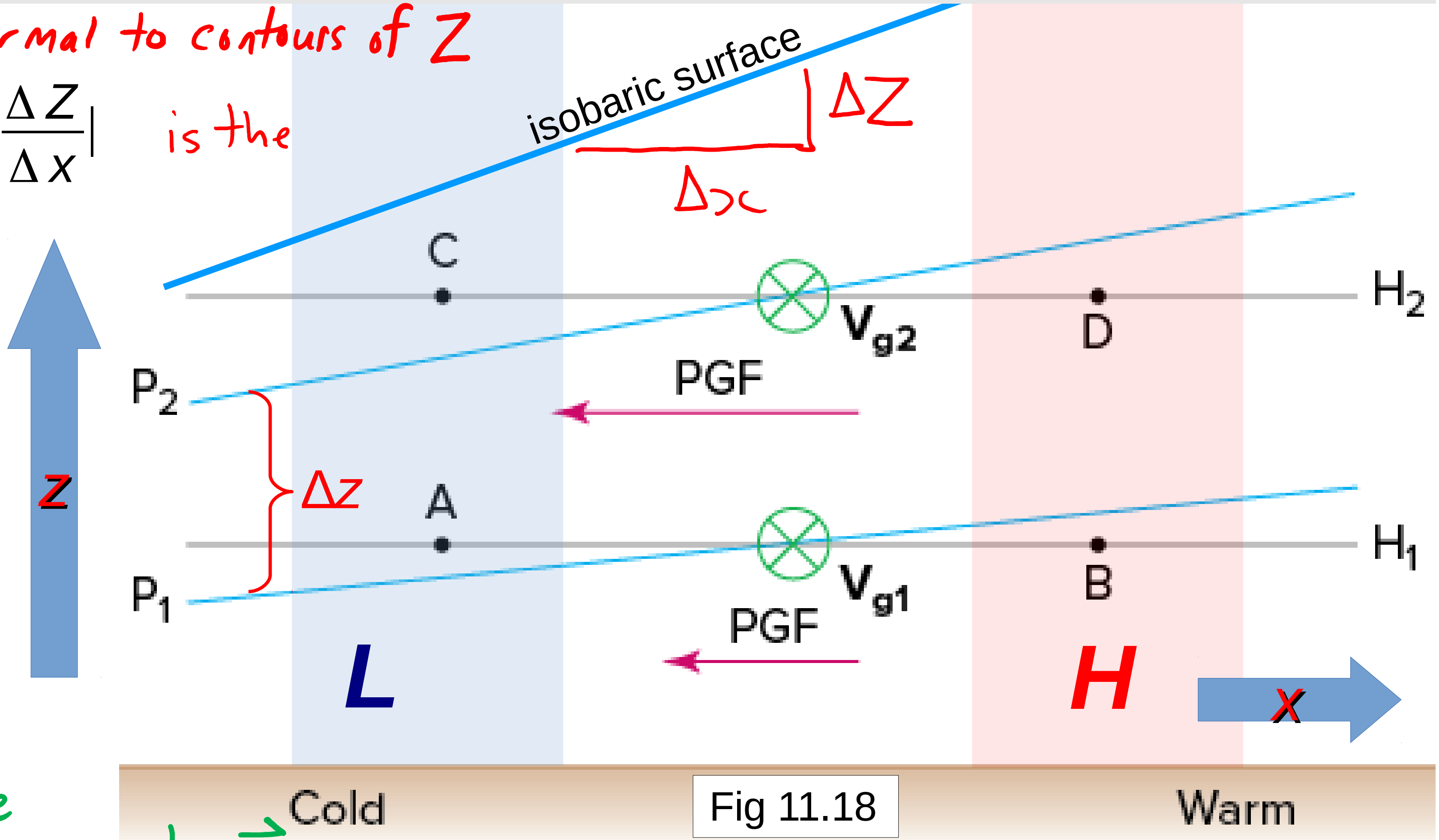


Fig 11.18

What do the green symbols denote? The geostrophic wind vector $\vec{V}_g$

- recall hypsometric eqn

$$\Delta z \equiv Z_2 - Z_1 = \left[\frac{R_d \; \overline{T}_v}{g} \right] \ln \frac{P_1}{P_2}$$

$$\left(\frac{\Delta P}{\Delta z} = -\rho g \right)$$

- successive isobaric surfaces have greater slope
- therefore V_g increases with increasing height

Still with x oriented perpendicular to isolines of height Z,

$$V_{g2} = \frac{g}{f_c}\left[\frac{\Delta Z}{\Delta x}\right]_2$$

$$V_{g1} = \frac{g}{f_c}\left[\frac{\Delta Z}{\Delta x}\right]_1$$

these are speeds, not vectors

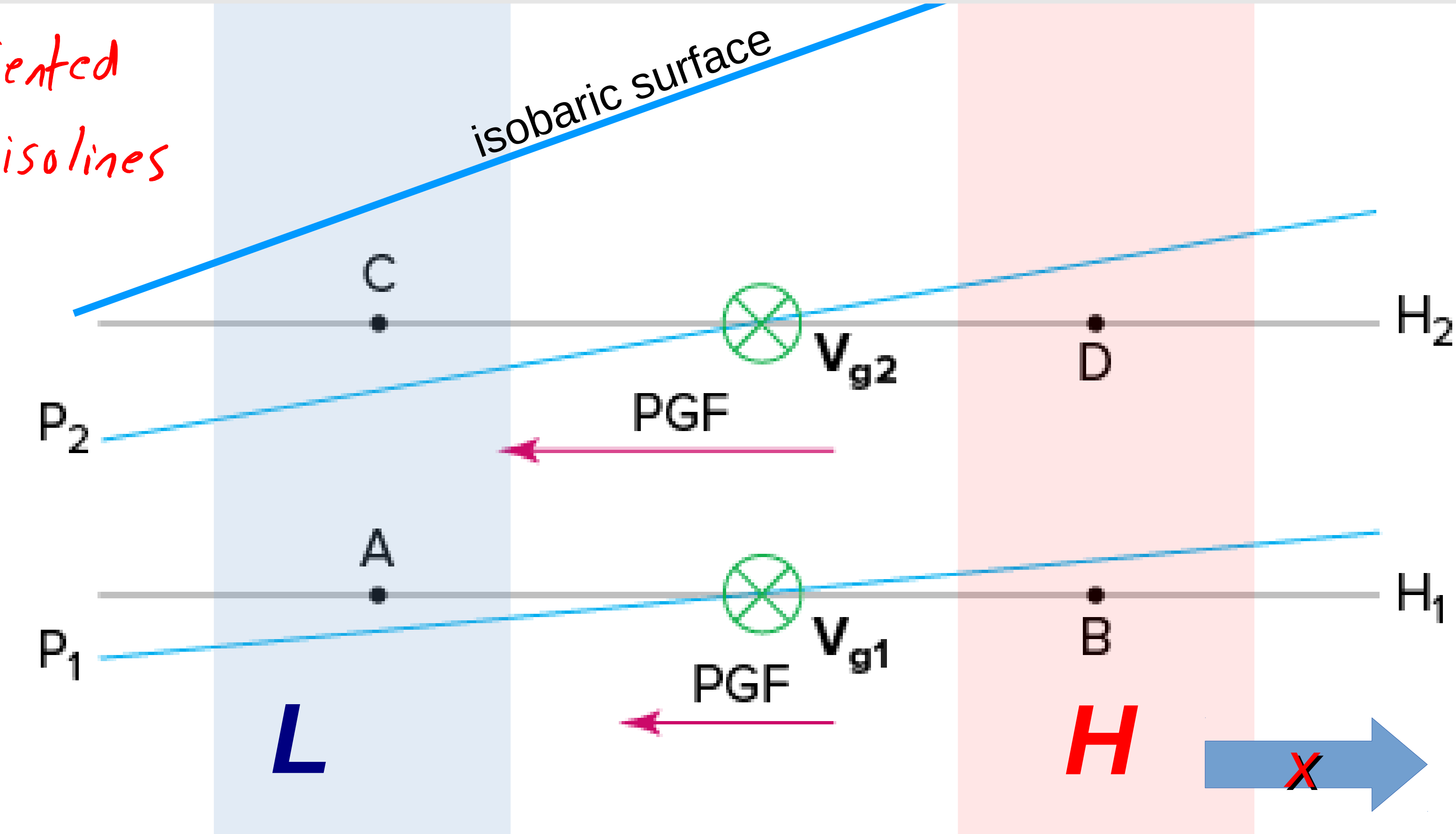


Fig 11.18

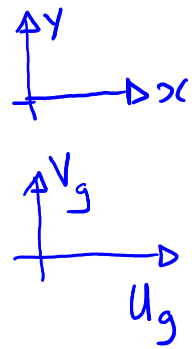
The "thermal wind" is the vector difference (sign + magnitude) between V_g at two levels...

$$\vec{V}_{T21} = \vec{V}_{g2} - \vec{V}_{g1} \quad \text{or} \quad \vec{V}_{g2} = \vec{V}_{g1} + \vec{V}_{T21}$$

Now let *x* point eastward and *y* point northward

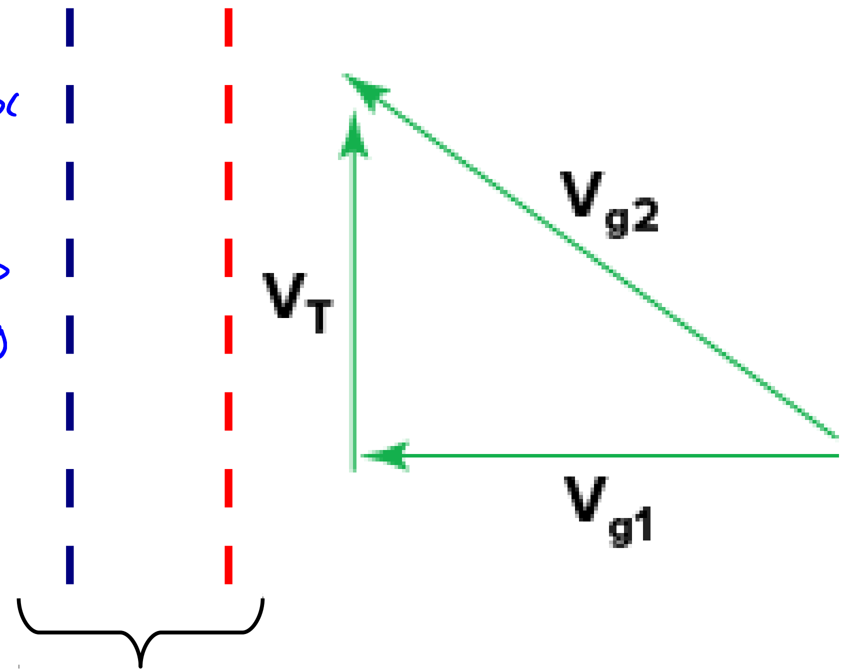
$$\vec{V}_{g2} = \underbrace{(U_{g2}, V_{g2})}_{\text{two components}} = \frac{g}{f_c}\left(-\frac{\Delta Z_2}{\Delta y}, \frac{\Delta Z_2}{\Delta x}\right)$$

$$\vec{V}_{g1} = (U_{g1}, V_{g1}) = \frac{g}{f_c}\left(-\frac{\Delta Z_1}{\Delta y}, \frac{\Delta Z_1}{\Delta x}\right)$$



$$\vec{V}_{T_{21}} = \vec{V}_{g2} - \vec{V}_{g1} = \frac{g}{f_c}\left(-\frac{\Delta(Z_2 - Z_1)}{\Delta y}, \frac{\Delta(Z_2 - Z_1)}{\Delta x}\right)$$

or if $h = Z_2 - Z_1$ ("thickness")



thickness contours

then

$$\vec{V}_{T_{21}} = \frac{g}{f_c}\left(-\frac{\Delta h}{\Delta y}, \frac{\Delta h}{\Delta x}\right)$$

For any layer "12" the thermal wind vector is parallel to the thickness lines (isolines of $Z_2 - Z_1$) for that layer and (in the N. hemisp.) blows with cold air on its LEFT

This is an example of a "veering" wind: can you see that the setup implies warm advection?

Wind vector at level 2 is the "resultant" of adding the thermal wind vector to the wind vector at level 1

$$\vec{V}_{g2} = \vec{V}_{T_{21}} + \vec{V}_{g1}$$

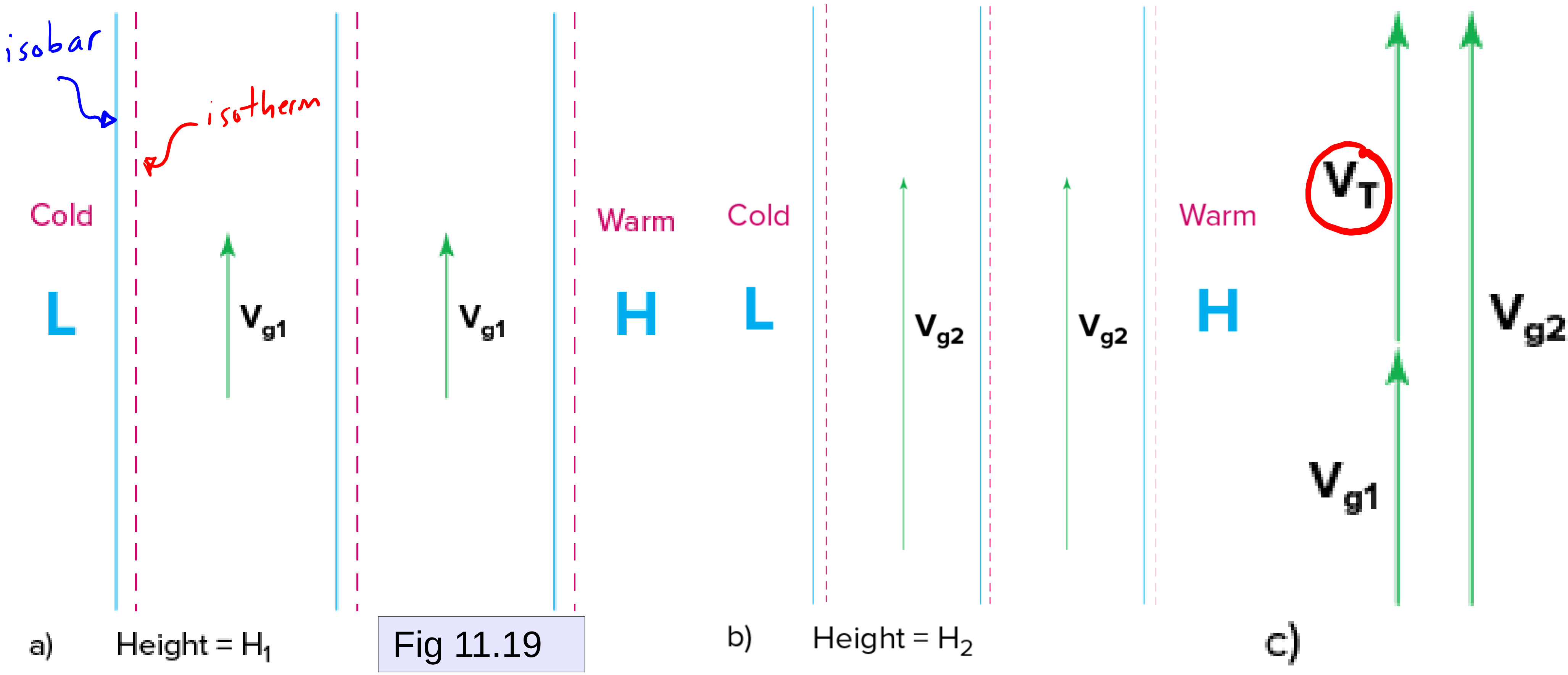


Fig 11.19

- thermal contours (isotherms, thickness isolines) are parallel to isobars and height contours
- thermal wind is parallel to geostrophic wind
- geostrophic wind *direction* not changing with height
- no thermal advection occurs because the wind runs // to isotherms

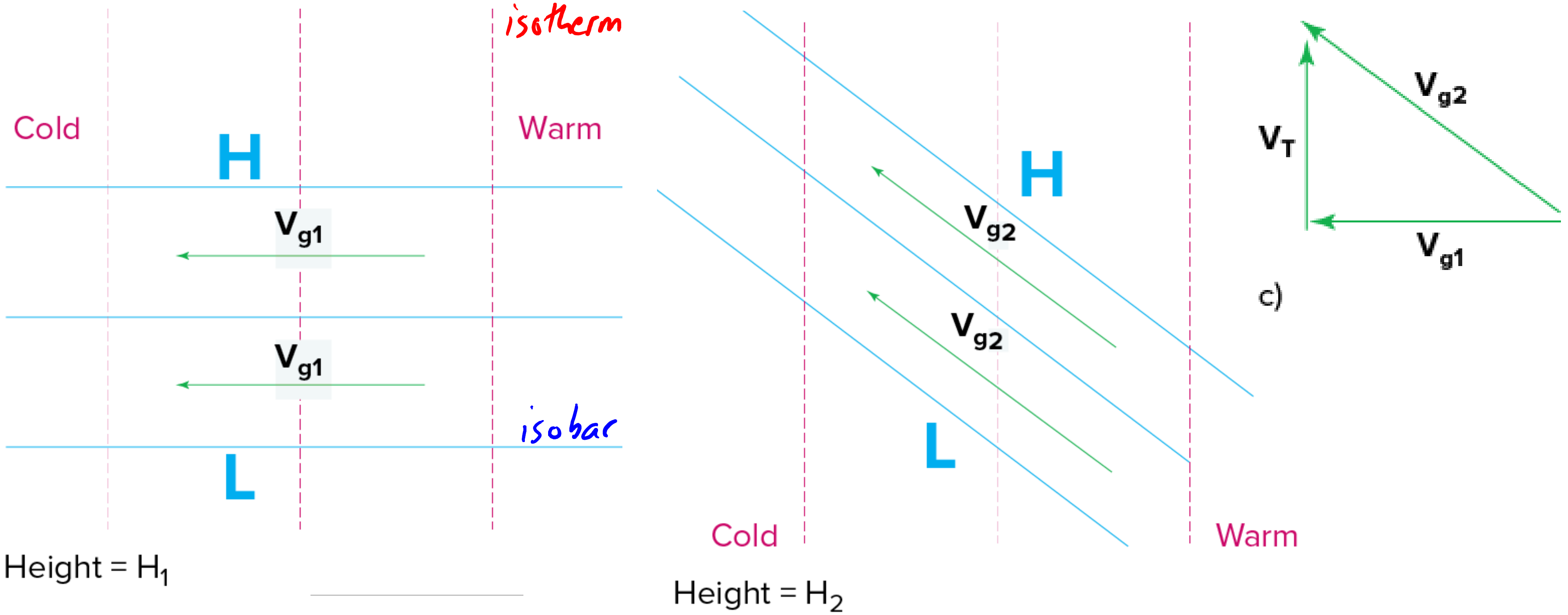
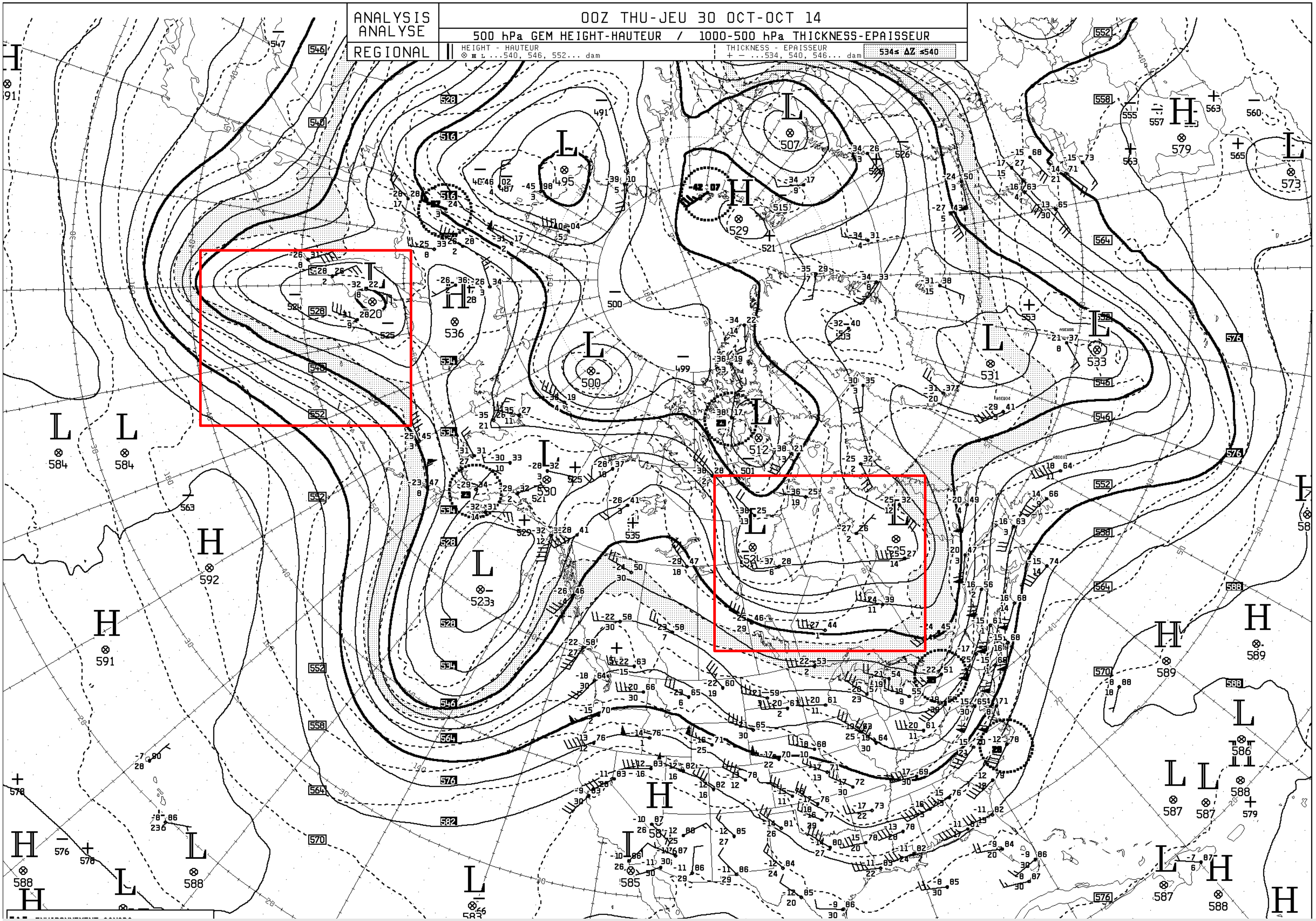
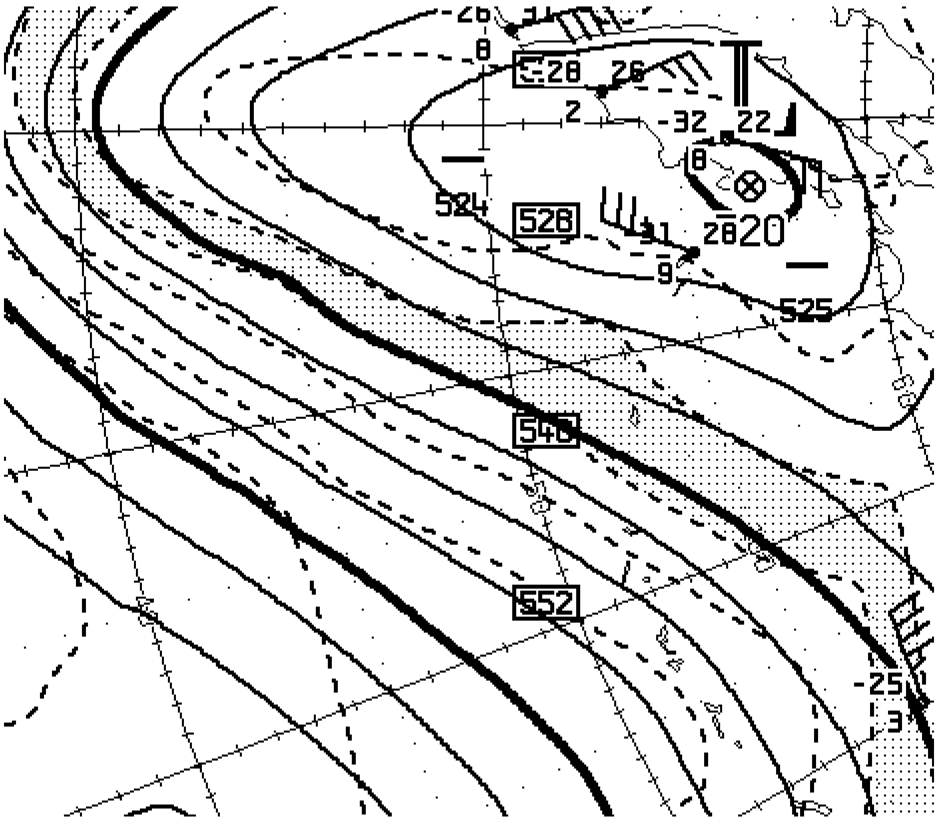


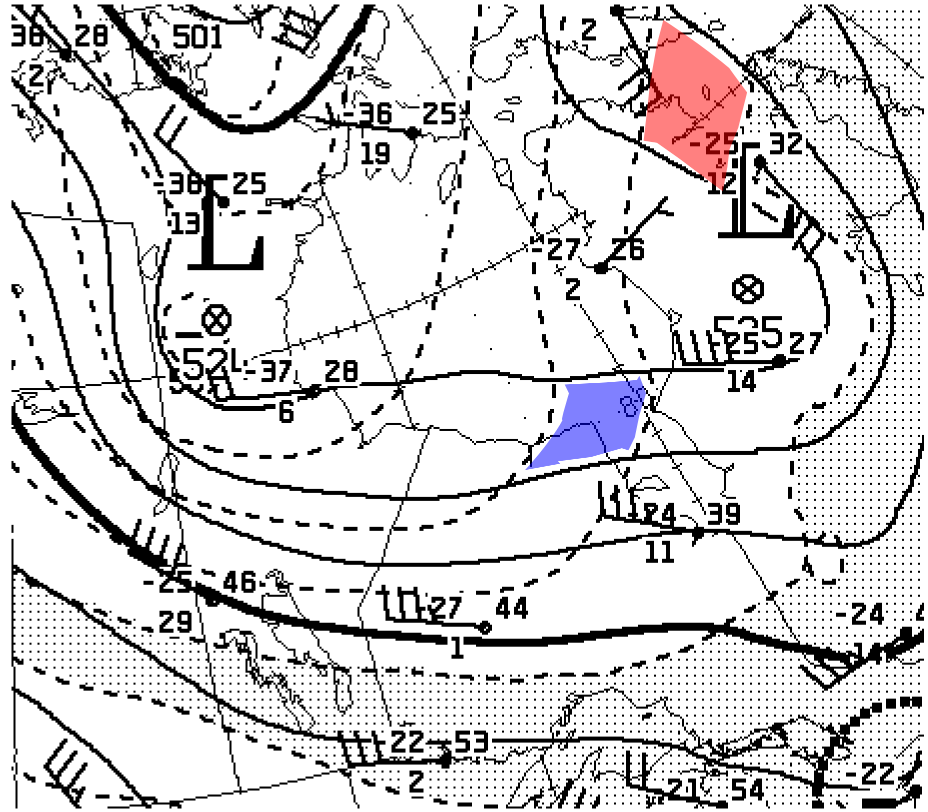
Fig 11.21

- isotherms and thickness contours are NOT parallel to isobars and/or height contours
- thermal wind NOT parallel to geostrophic wind
- geostrophic wind *direction* (and speed) changing with height
- thermal advection occurring



CMC 500 hPa analysis, 00Z 30 Oct. 2014 – next page, a zoomed-in view





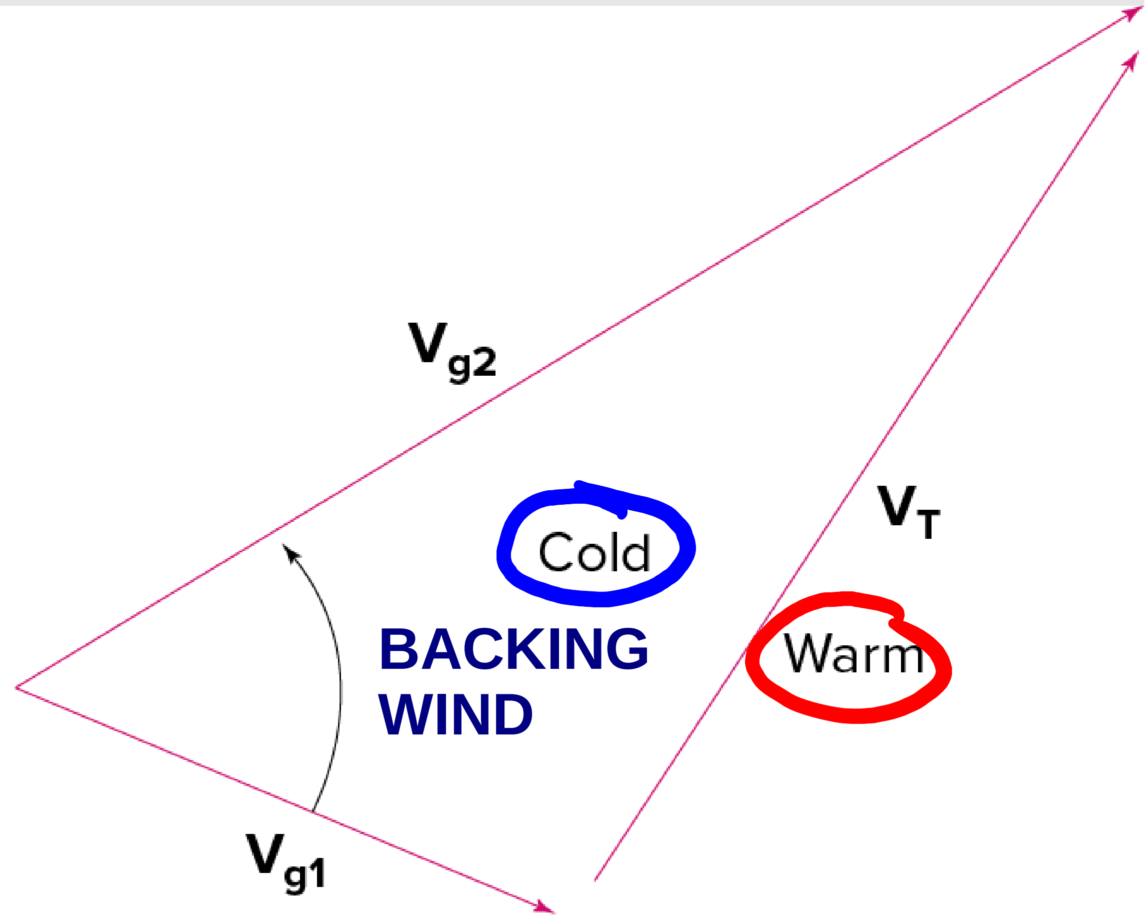
- barotropic – no thickness advection

- baroclinic – thickness advection

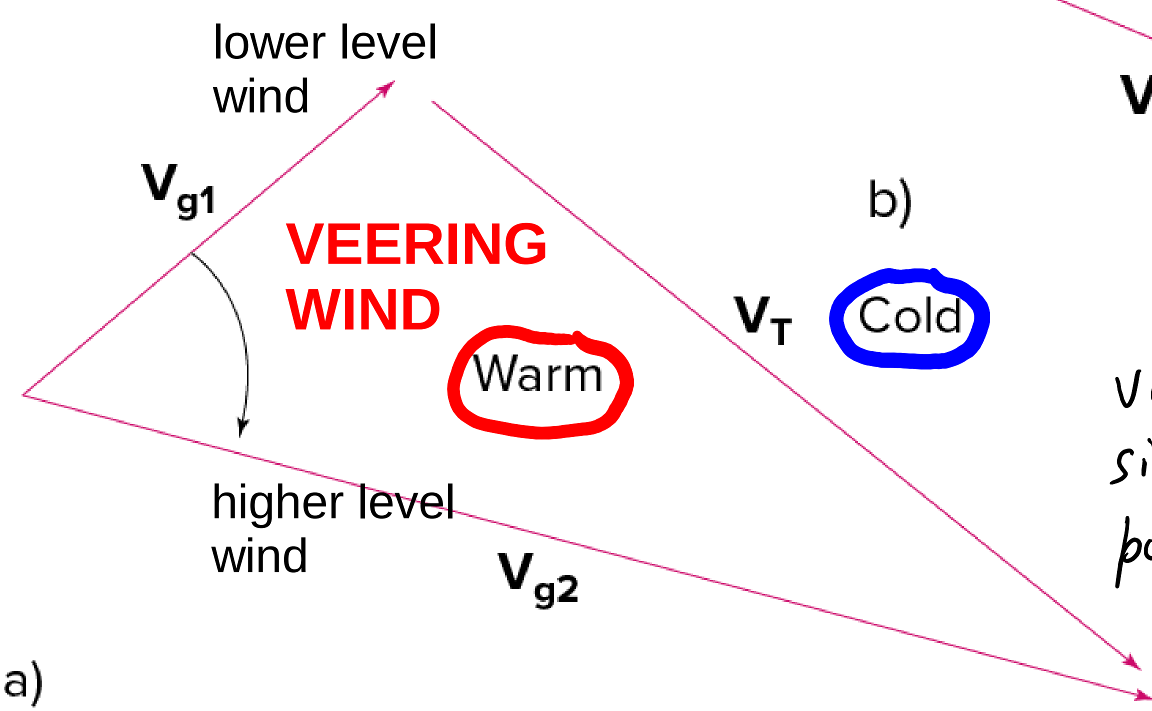
Veering wind = warm advection

Backing wind = cold advection

Remember these direction changes are above the ABL – the mechanism has nothing to do with friction, it is related to the spatial pattern of temperature (equivalently, thickness, or the "thermal field")



b)



a)

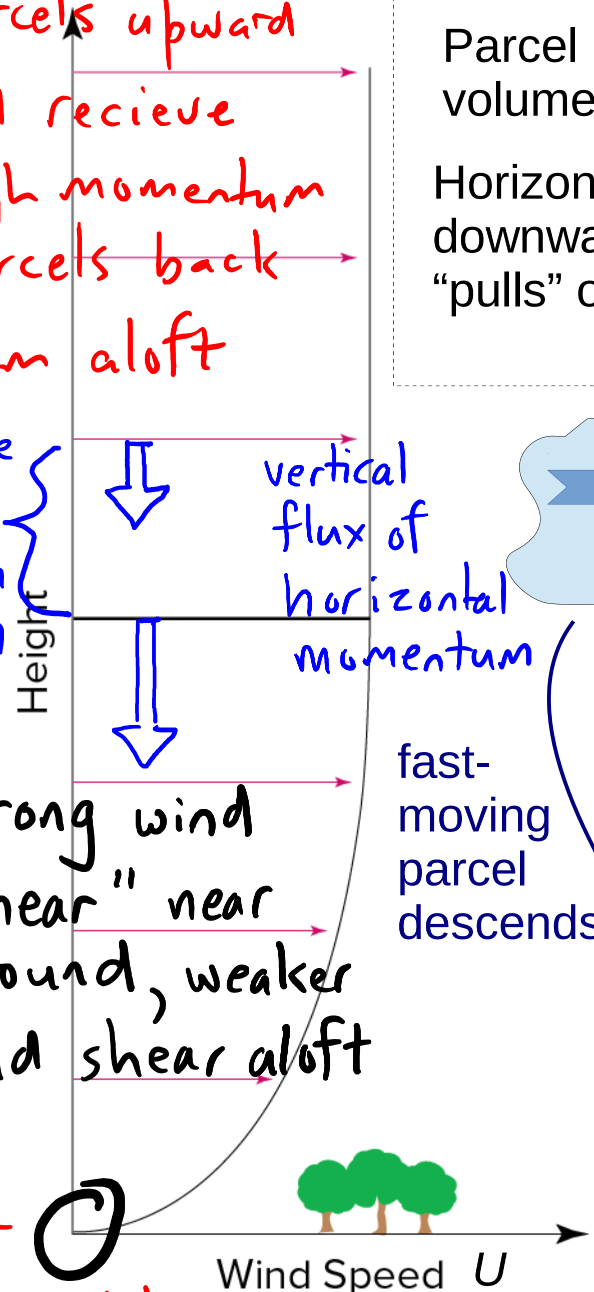
By inspection of the winds' variation with height from a single radiosonde, we can infer possibility for thermal advection (and the whereabouts of warmer and colder air)

Fig 11.24

Near surface layers export low momentum parcels upward and recieve high momentum parcels back from aloft

more loss than gain

vertical flux of horizontal momentum



strong wind "shear" near ground, weaker wind shear aloft

"no slip condition"

Parcel's velocity (u,v,w)

Parcel momentum per unit volume: ρ (u, v, w)

Horizontal mtm transported downward by eddies and "pulls" on the surface

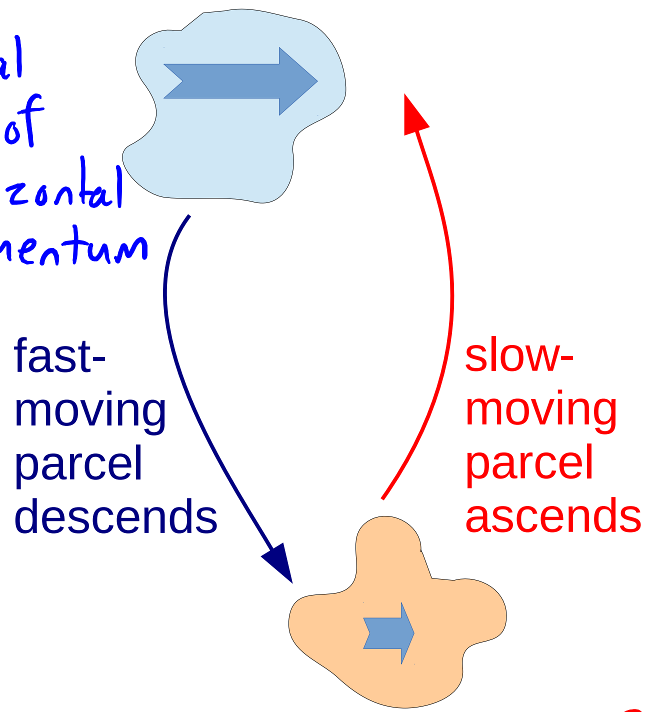


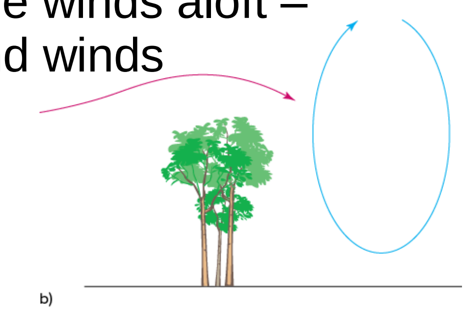
Fig 11.7 (modified)

- effects of surface friction carried upward by turbulent eddies
- the greater the level of turbulence, the stronger the vertical exchange rate (of heat, momentum, moisture...)

turbulence increases with increasing surface roughness, increasing wind speed and increasing thermal instability ($Q_H > 0$)

- strong daytime turbulence brings down fast-moving (high momentum) parcels from near the top of the PBL and these push along the air near ground – strong surface winds result
- weak night-time turbulence (stable stratification) means weaker "coupling" of surface air to the winds aloft – slower near-ground winds

STOPPED HERE MON 21 NOV.



PGF is indep. of height (in simplest case)

PGF not changing with height

Wind speed decreases near ground due to friction force F, decreasing CF

New balance of three forces

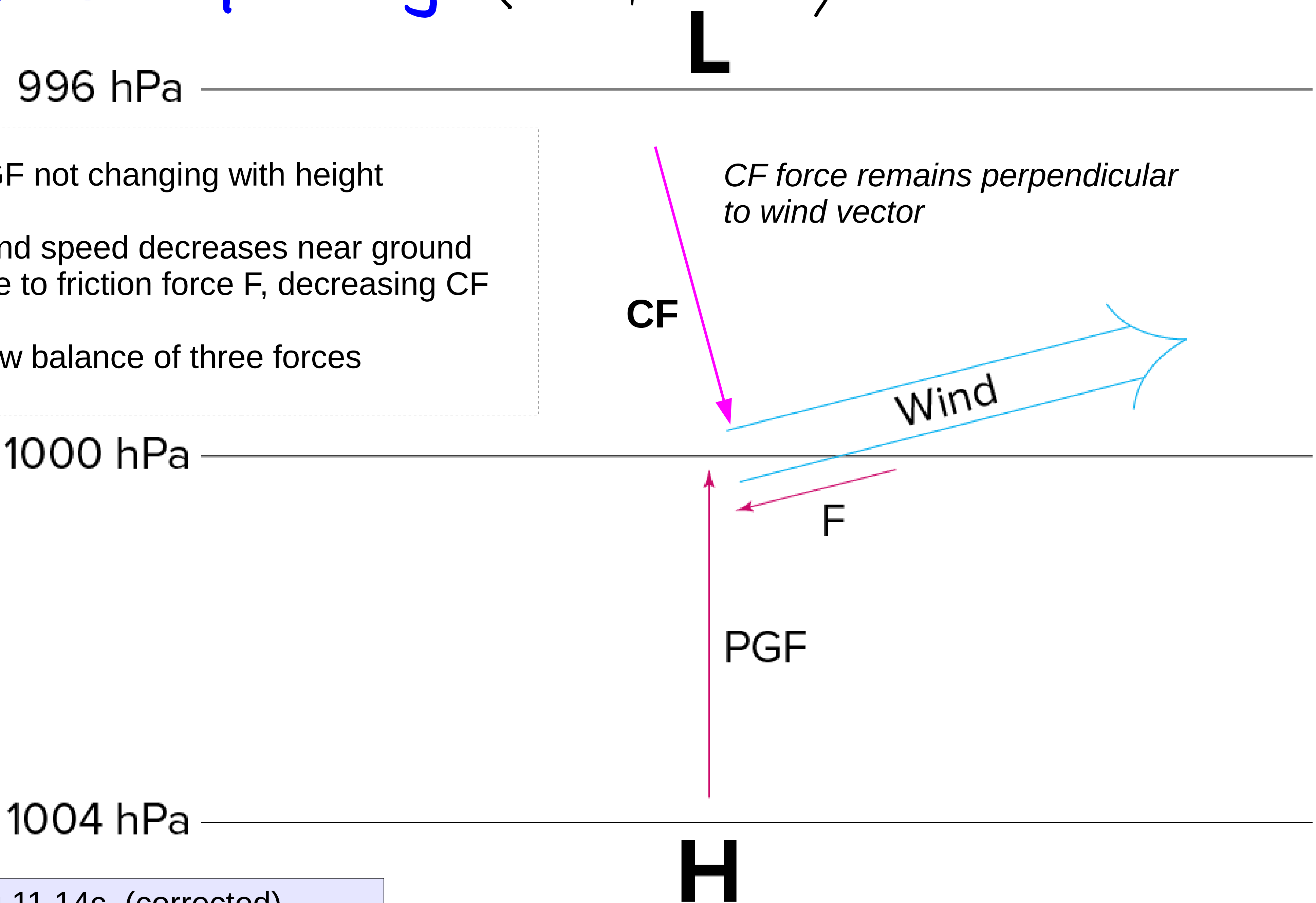


Fig 11.14c (corrected)

If upper level divergence exceeds surface convergence, the sfc low will deepen

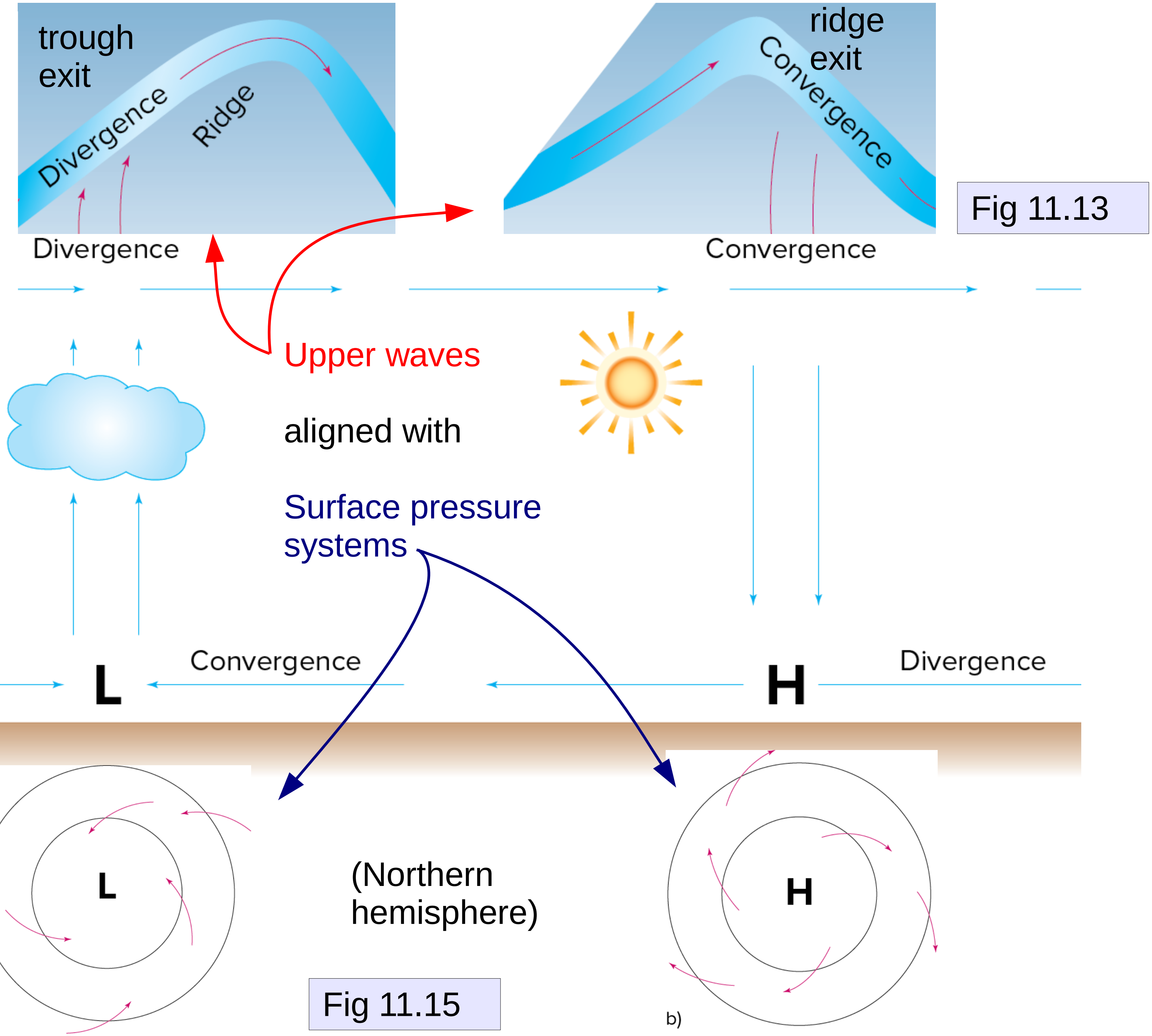


Fig 11.13

Fig 11.17

Fig 11.15

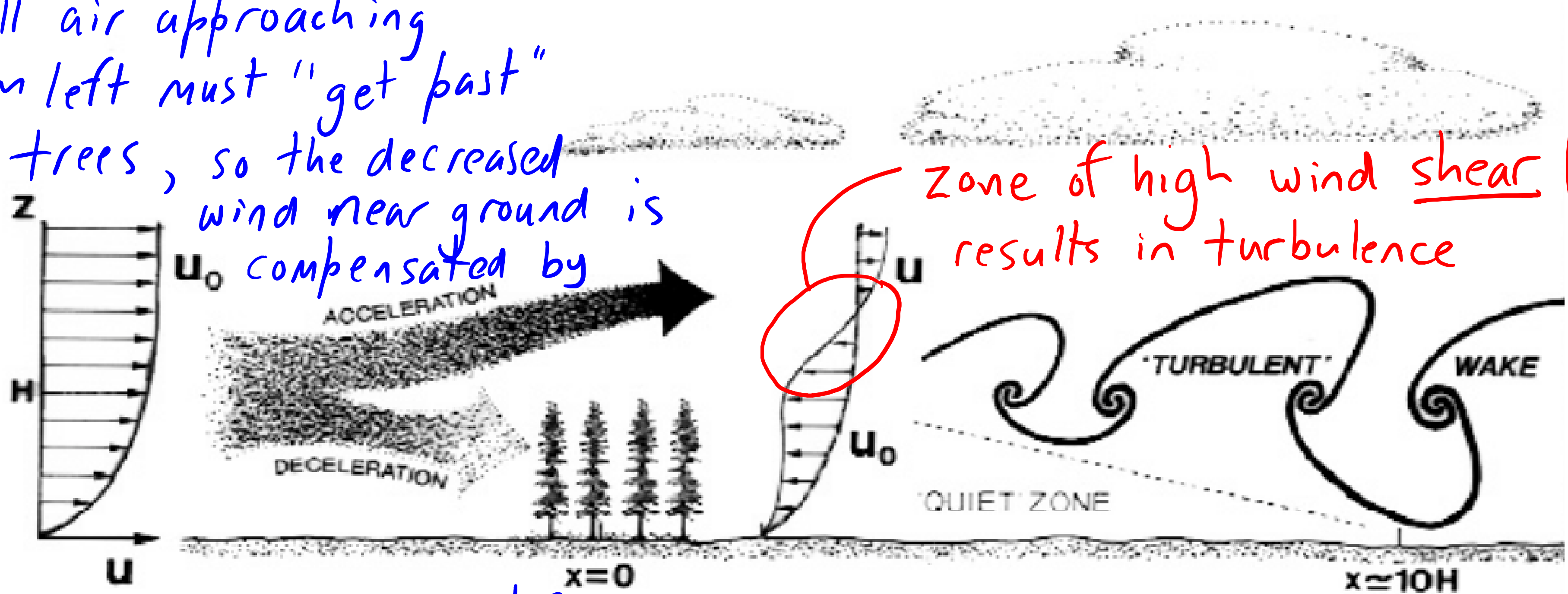
Circulation patterns on scales smaller than the synoptic scale that we view on the weather charts. These are "shallow" flows, typically taking place within the ABL

- micrometeorological flows
- mesoscale flows

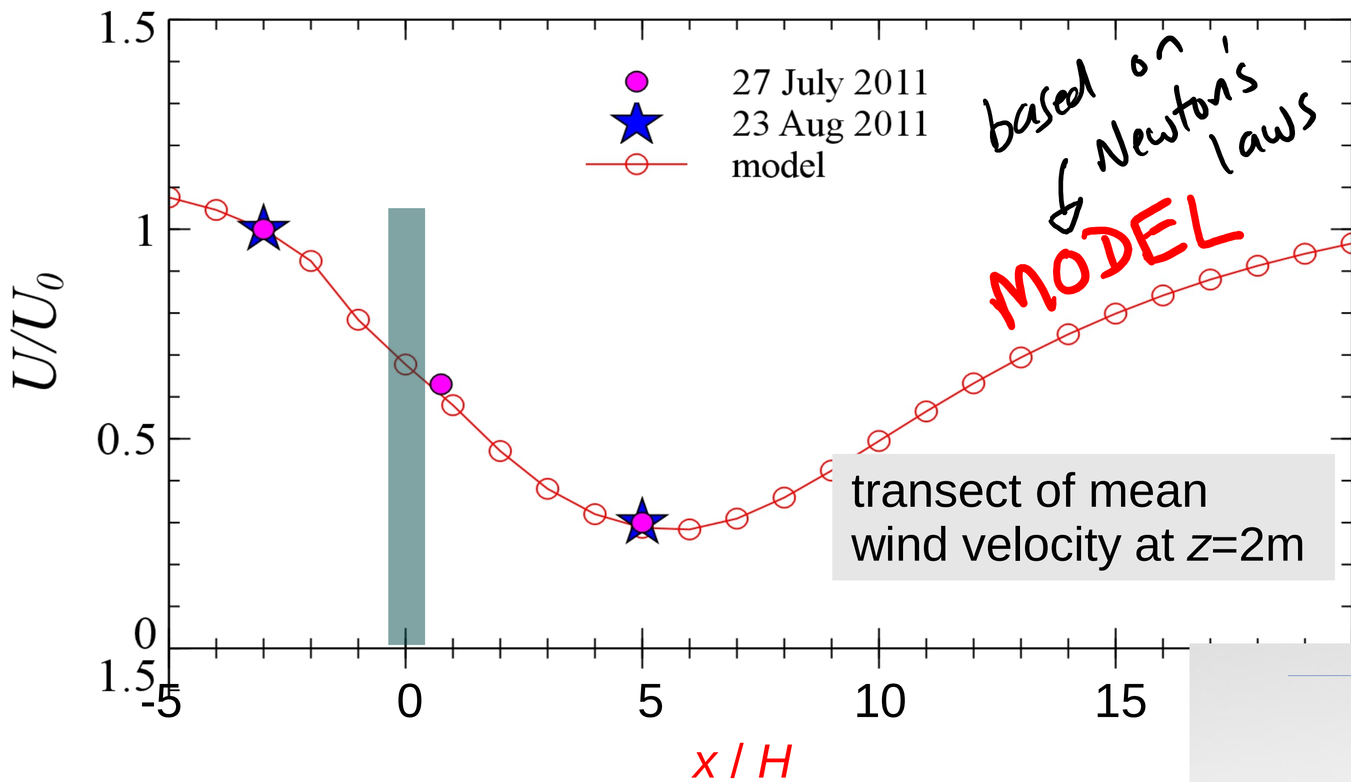
These flows may be driven by surface temperature and/or moisture gradients, by topography, by changes in the surface elevation and/or cover, by obstacles,...

In micro- and meso-scale meteorology we describe properties in the PBL in terms of their *statistics* (average=mean, standard deviation,...), defined over a sampling period of ~ 15 to 60 min

All air approaching from left must "get past" the trees, so the decreased wind near ground is compensated by an increased wind aloft.

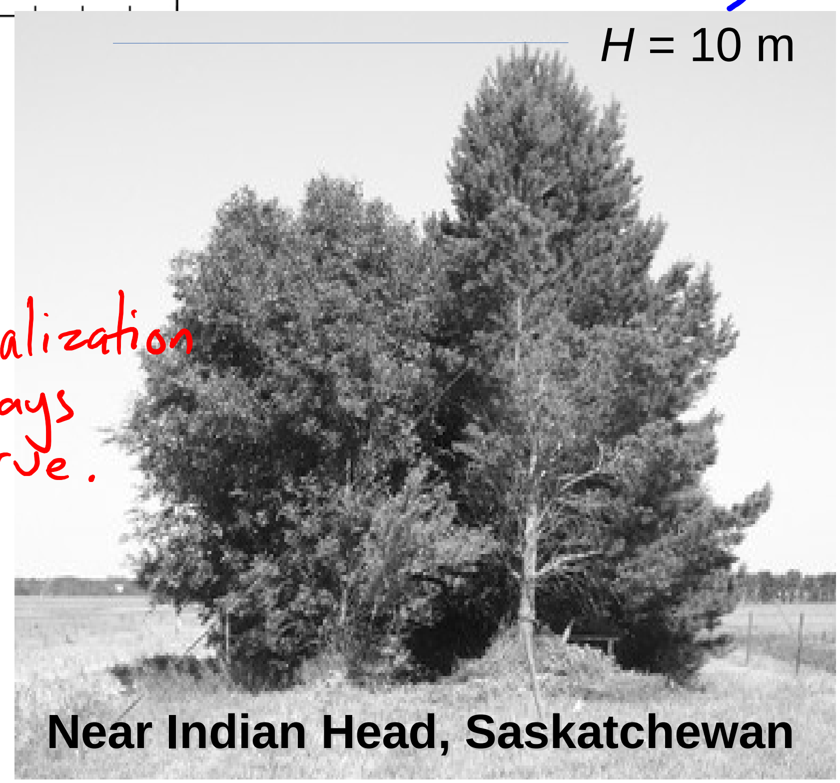


Zone of high wind shear $\left(\frac{\Delta u}{\Delta z}\right)$ results in turbulence



U_0 the time average speed measured upwind from the shelterbelt at 2 m AGL, and U the time average speed elsewhere at the same height 2 m AGL

(Measured when wind dir. ⊥r to windbreak)



Near Indian Head, Saskatchewan

Notice that variables have been "normalized", i.e. made dimensionless by choosing "scales" for length and velocity

Only because of the normalization do results from different days fall on the same curve.

50 m downwind of the trees the wind speed was reduced to only 30% of its value in the open (upwind)

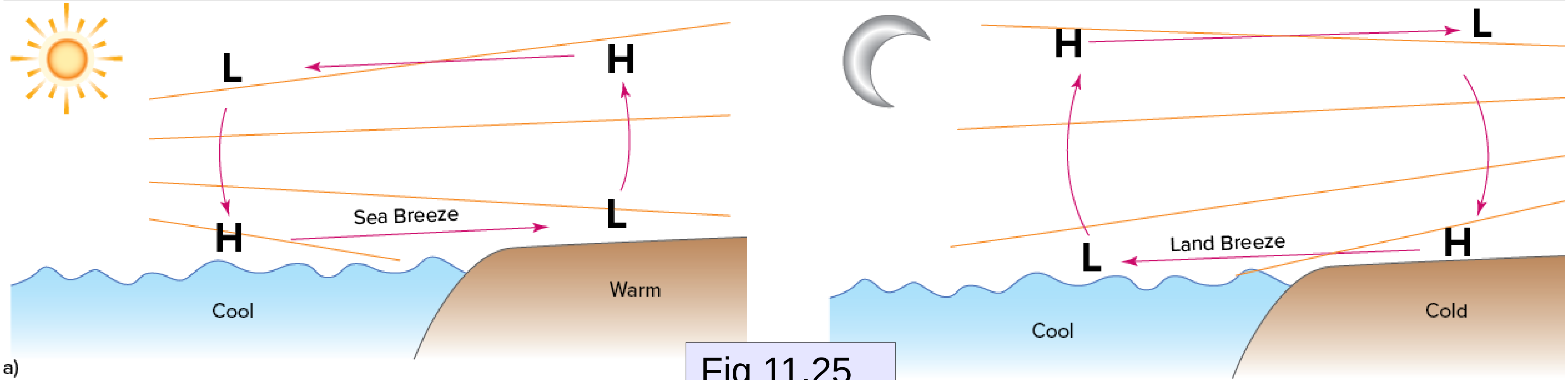
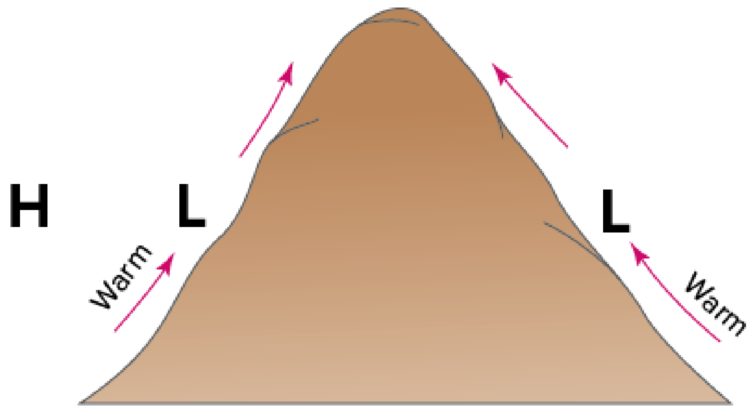


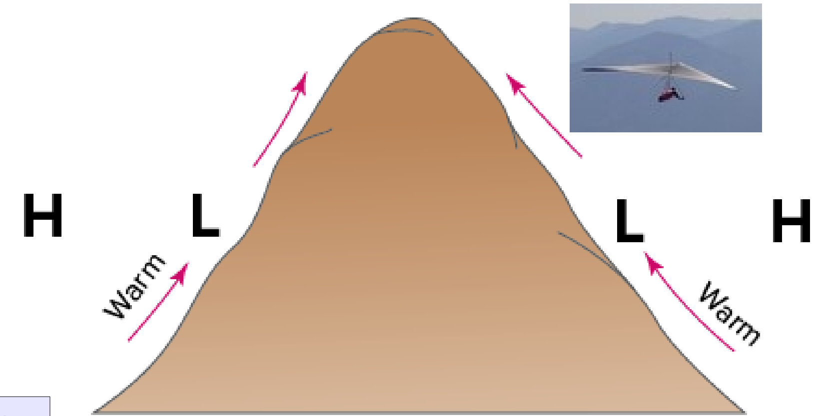
Fig 11.25

- driven by surface thermal gradient originating in different energy balances over land and sea (and thermal inertia of the ocean) $Q^* = Q_H + Q_E + \Delta Q_S$
- results in a "thermal" pressure system
- shallow circulation (~ 1km deep) may penetrate up to 100 km inland
- masked by strong synoptic scale winds
- stronger on days with strong solar heating, and strongest mid afternoon when maximal ΔT
- sea breeze usually stronger than land breeze, as the nightime ΔT tends to be smaller

Question: do you think the Coriolis force affects these flows? Why?

YES. Lifetime is of order $1/f_c$





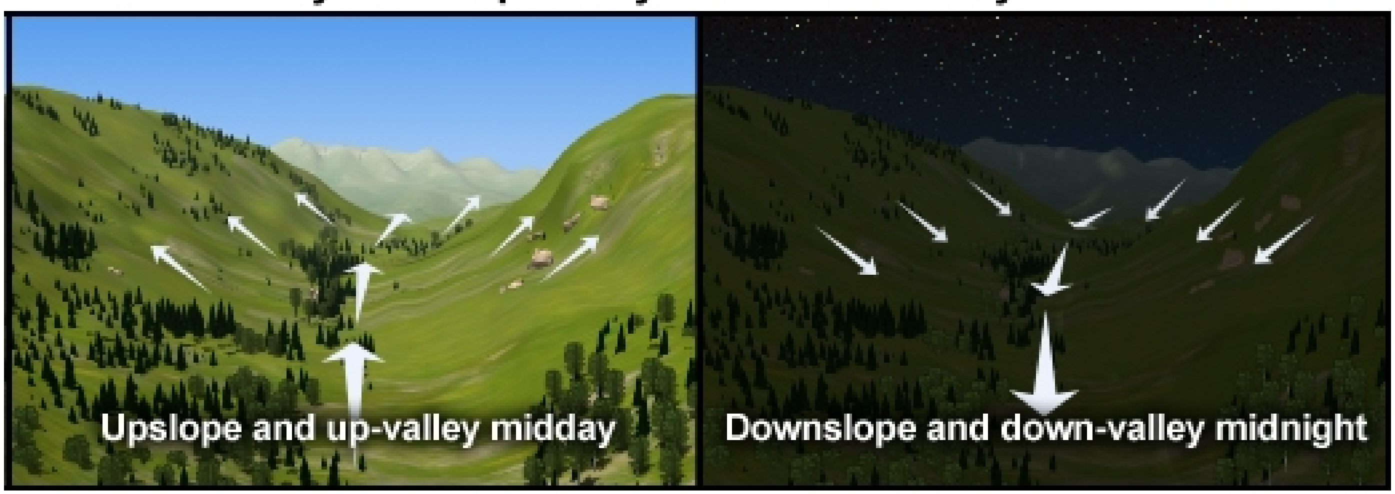
a)

Fig 11.29

- air in immediate contact with the slope is warmed (or cooled) by the surface heat flux Q_H, and this results in a buoyancy force

$$F_B = g \frac{T_{\text{slope}} - T_{\text{mid-valley}}}{T_{\text{mid-valley}}}$$

Diurnal Cycle of Up-valley and Down-valley Circulations



Topics/concepts covered

- geostrophic wind spd. and dir. may vary with height – due to the thermal pattern
- thermal wind vector not an actual wind, but *difference* of wind vectors at 2 levels
- it is oriented parallel to thickness lines with cold air on the left (in N. hemi)
- a veering (backing) wind is associated with warm (cold) advection
- mechanism of friction (vertical transport of momentum by turbulent eddies)
- factors enhancing turbulence in the Atmospheric Boundary Layer (ABL)
- friction slows and turns the wind in the ABL
- surface convergence in lows, surface divergence in highs
- coordination of surface convergence with divergence in trough exit aloft makes for a "well supported" storm – strong ascent
- necessity to adopt statistical description of properties in the ABL
- basis for dichotomy between “disturbed” and “undisturbed” flows
- “normalizing” variables by adopting suitable “scales”
- sea breeze – shallow mesoscale thermal pressure system
- slope winds (gravity-driven flows)