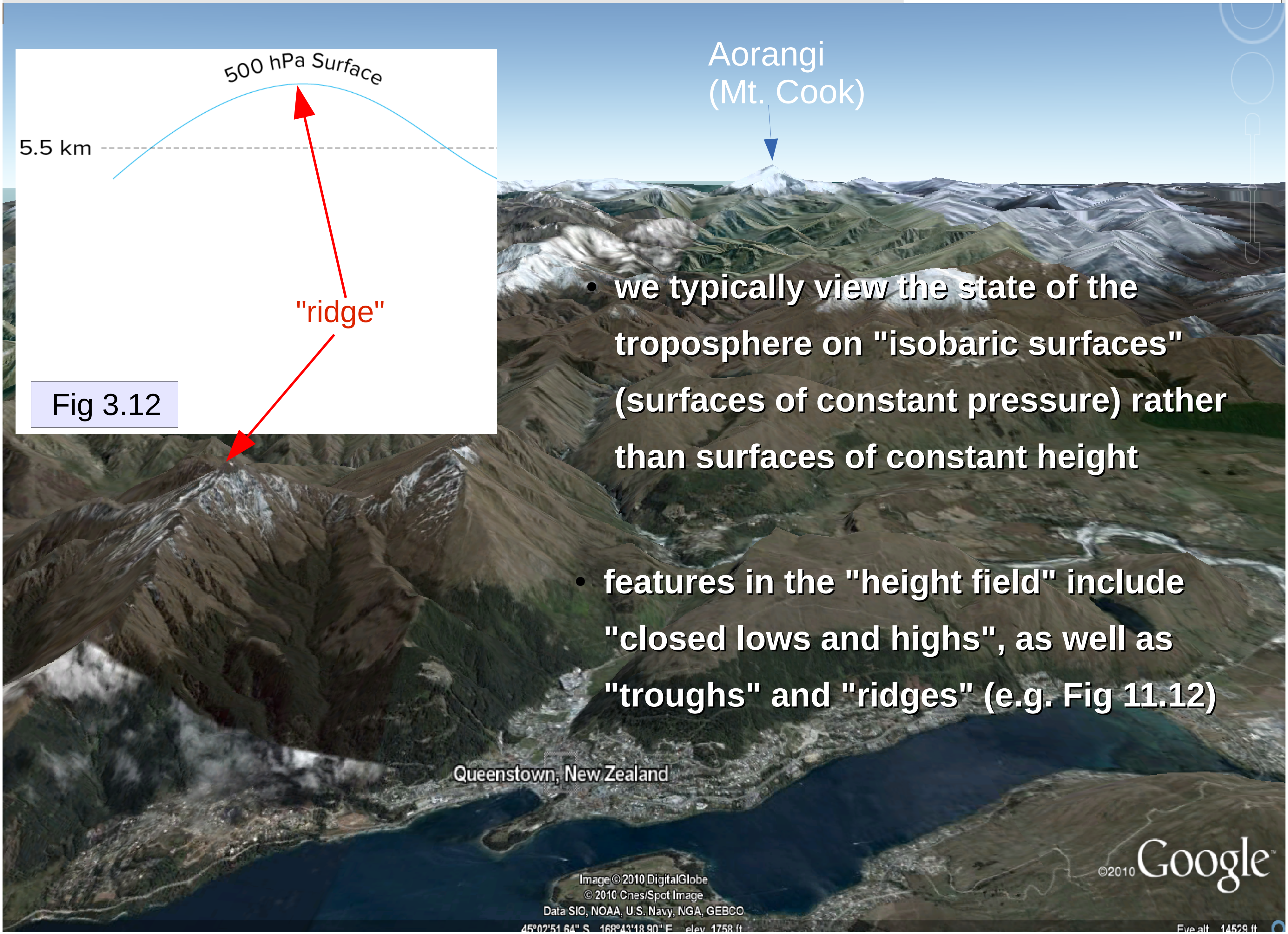


Fig 3.12

- **we typically view the state of the troposphere on "isobaric surfaces" (surfaces of constant pressure) rather than surfaces of constant height**

- **features in the "height field" include "closed lows and highs", as well as "troughs" and "ridges" (e.g. Fig 11.12)**

- wind "follows" height contours
- upper lows tend to be better defined than the highs
- weak gradients at low latitudes
- meandering airstreams (waves in the upper flow)

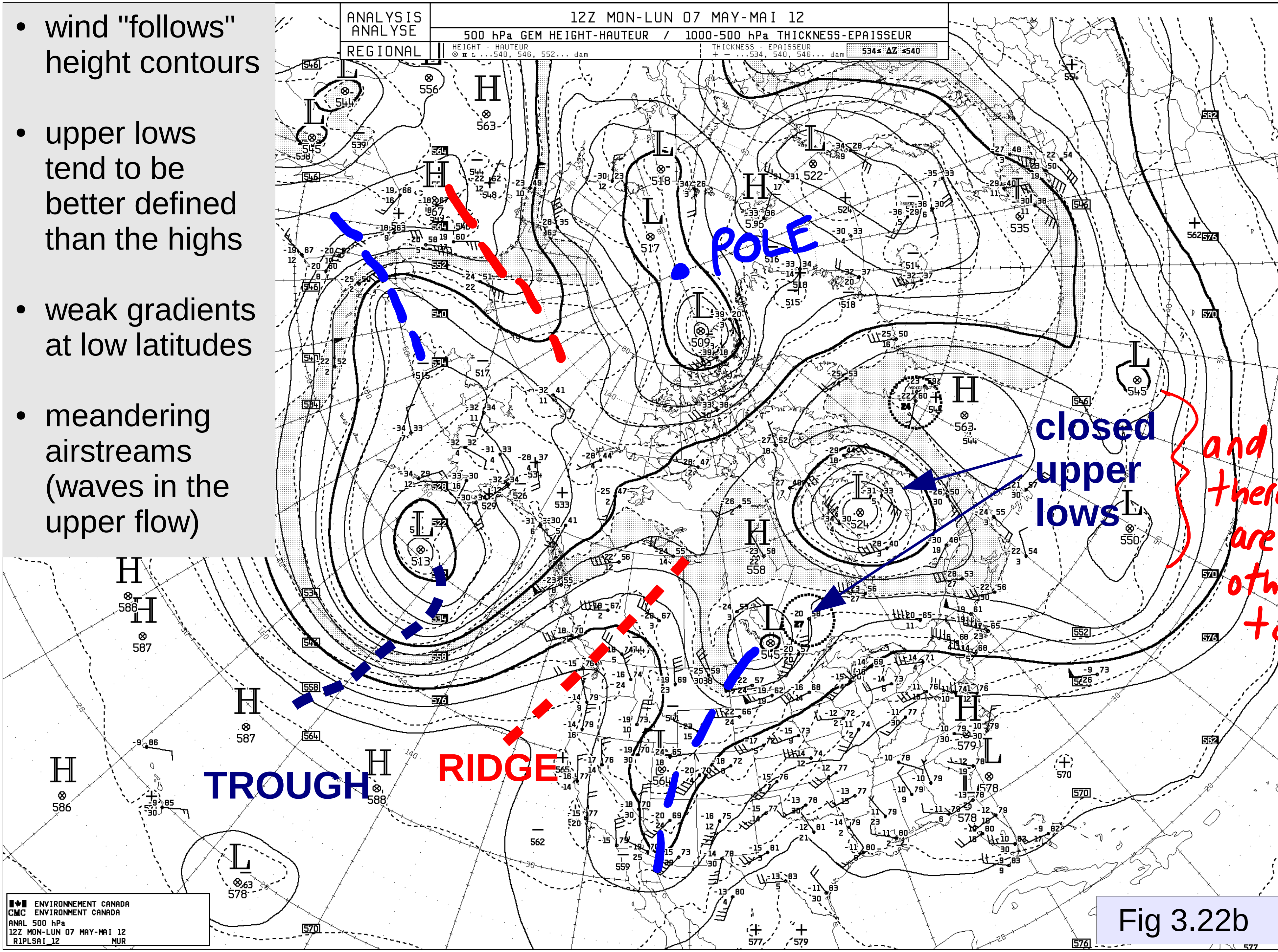
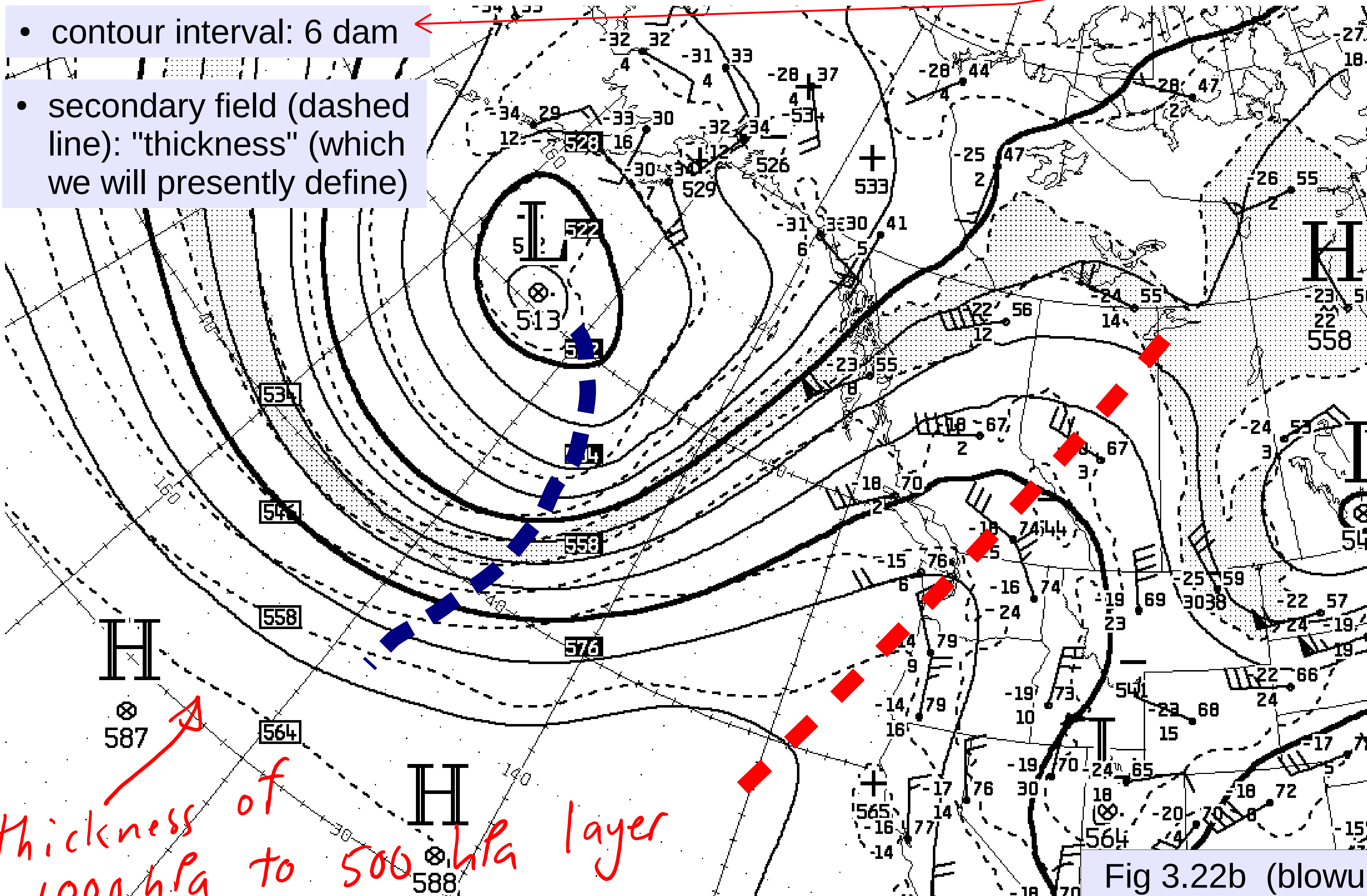


Fig 3.22b

- as you move perpendicularly away from the trough axis, heights rise
- as you move perpendicularly away from the ridge axis, heights fall

The height contour interval is 6 dam for the 850, 700 and 500 hPa surfaces; but it is 12 dam for the 250 hPa surface.

- contour interval: 6 dam
- secondary field (dashed line): "thickness" (which we will presently define)

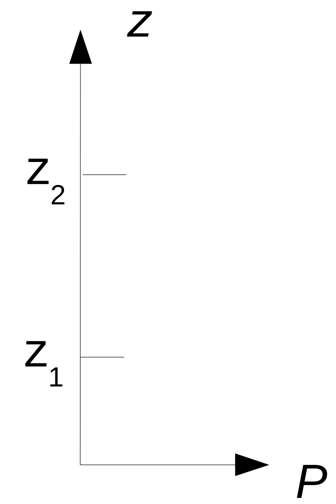


thickness of 1000 hPa to 500 hPa layer

Fig 3.22b (blowup)

$$P = \rho \ R_d \ T_v \qquad\qquad \frac{\Delta P}{\Delta z} = -\ \rho \ g$$

eliminate density



$$\frac{dP}{dz} = -\left(\frac{P}{R_d \ T_v}\right) g \qquad \text{(Eqn 3.10)}$$

ρ

"the (rate of) change in pressure with height is greater in cold air than it is in warm air"

Note: rather than use the "finite difference notation" Δ here, I have chosen to use the derivative symbol *d*. This is because I want to express the hydrostatic law in a form that applies at each and every point, whereas if the height increment were finite (denoted Δz rather than dz) I'd need T_v to be an average over the interval. This distinction is made here only to avoid any discomfort for those needing the calculus to be rigorous.

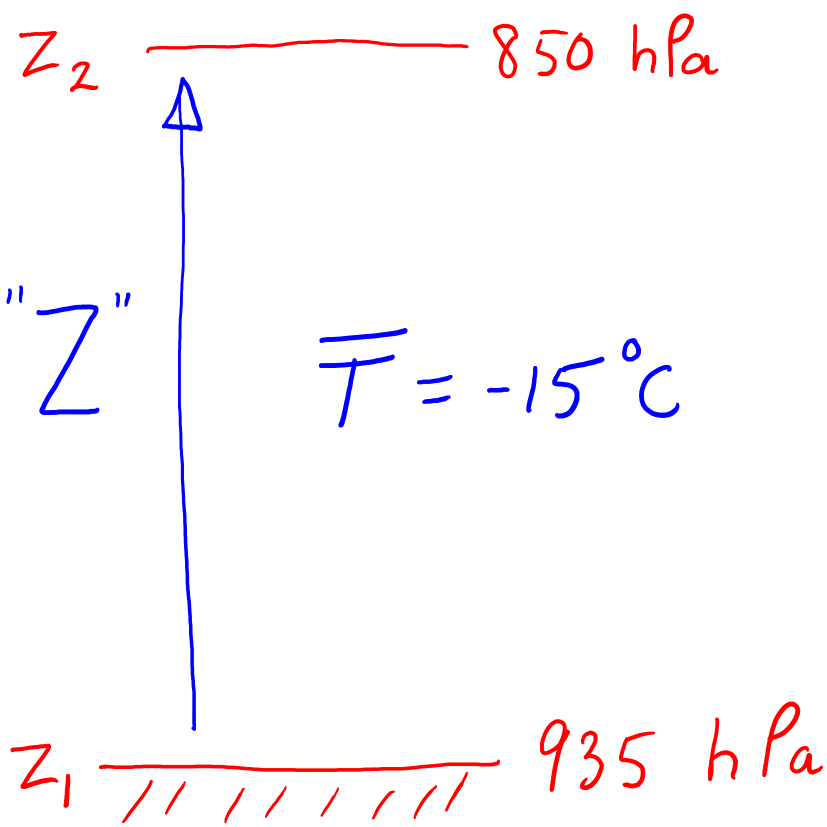
We now integrate upwards from height z_1 to height z_2, using the first mean value theorem of calculus (steps not shown):

$$\Delta z \equiv z_2 - z_1 = \left[\frac{R_d \ \overline{T}_v}{g}\right] \ln \frac{P_1}{P_2} \qquad \text{(Eq 3.11)}$$

Thickness

"the thickness of an atmospheric layer is the difference in height between two pressure surfaces" (that bound it). It is controlled by the layer's average (virtual) temperature

Suppose that at 00 UTC in Edmonton** on a certain day the pressure at ground level was P=935 hPa and that the average value of the temperature below the 850 hPa isobaric surface was -15°C: compute the height above ground of the 850 hPa surface (neglect the distinction between T and T_v, for at these temperatures the mixing ratio r is very small).



$$\Delta z \equiv z_2 - z_1 = \left[\frac{R_d\ \bar{T}_v}{g}\right] \ln\frac{P_1}{P_2}$$

$R_d = 287\ \mathrm{J\ kg^{-1}\ K^{-1}}$

$$\Delta z = \frac{287 \times (273.15 - 15)}{9.81} \ln\frac{935}{850}$$

$$\simeq 720\ \mathrm{m}$$

**Edmonton is about 700 m ASL (700 m above sea level), so surface pressures are much lower than the nominal 1000 hPa (=100 kPa = 10^5 Pa) figure for pressure at sea level

Let P_1 = 850 hPa, P_2 = 700 hPa. Suppose that layer mean T_v = -15°C at A and layer mean T_v = 5°C at B. What is the 850-700 hPa *thickness* at A and at B?

$$\Delta z = \frac{R_d}{g} \overline{T}_v \ln \frac{850}{700} = \frac{287}{9.81} \overline{T}_v \ln 1.214 = 5.68 \, \overline{T}_v$$

(i) at A, $\Delta z = 1466$ m (ii)

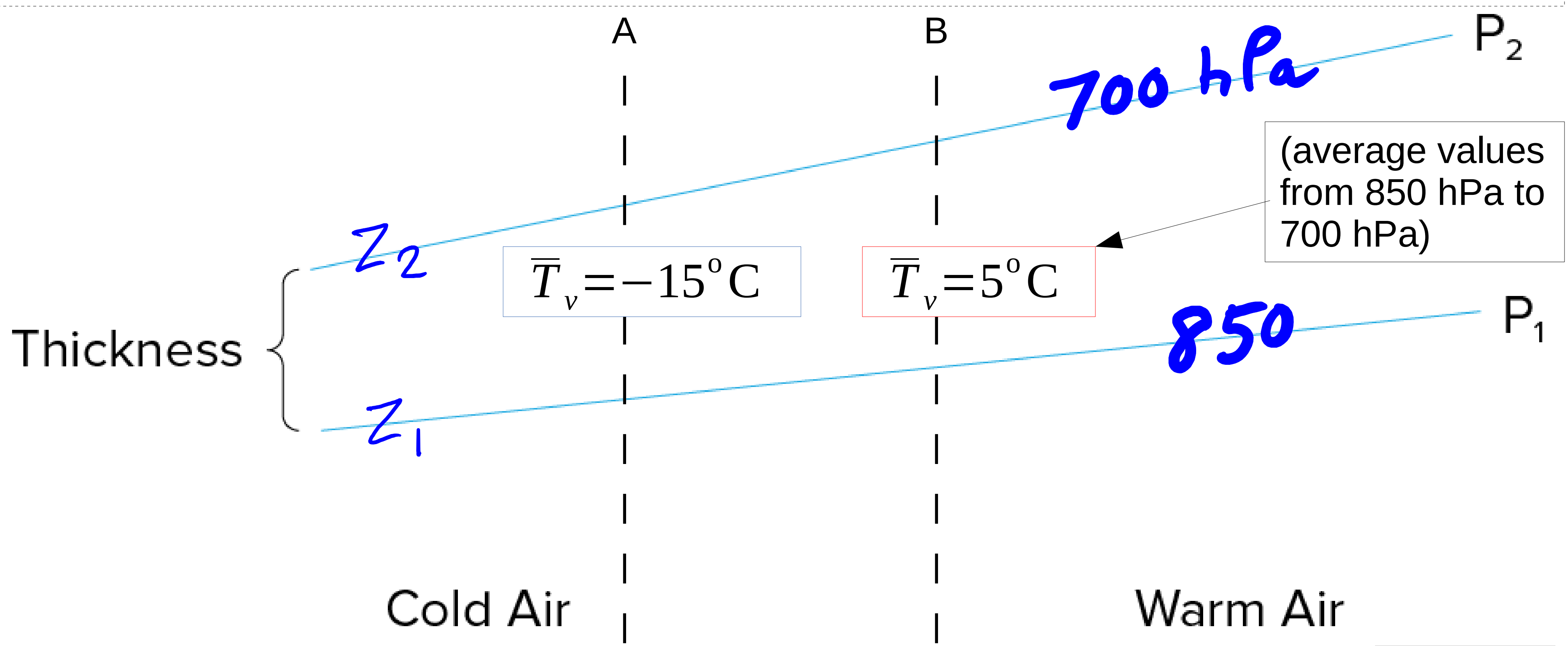


Fig 3.13

Again supposing that at 00 UTC in Edmonton on a certain day the pressure at ground level was P=935 hPa, compute sea level corrected surface pressure (MSLP) P_0, adopting a value of -15°C for the temperature of the (fictitious) air column down to sea level:

$$700\,m = \Delta z = Z_{EDM} - 0 = \frac{R_d}{g}\,\overline{T_v}\,\ln P_0/P$$

$$= \frac{287}{9.81}\,(273.15 - 15)\,\ln P_0/935$$

$$= 7552\,\ln\frac{P_0}{935}$$

$$\boxed{e^{\ln(x)} = x}$$

$$\ln\frac{P_0}{935} = 0.09269$$

$$\frac{P_0}{935} = e^{0.09269} = 1.097$$

$$\therefore\ P_0 = 1.097 \times 935 = 1026\ hPa$$

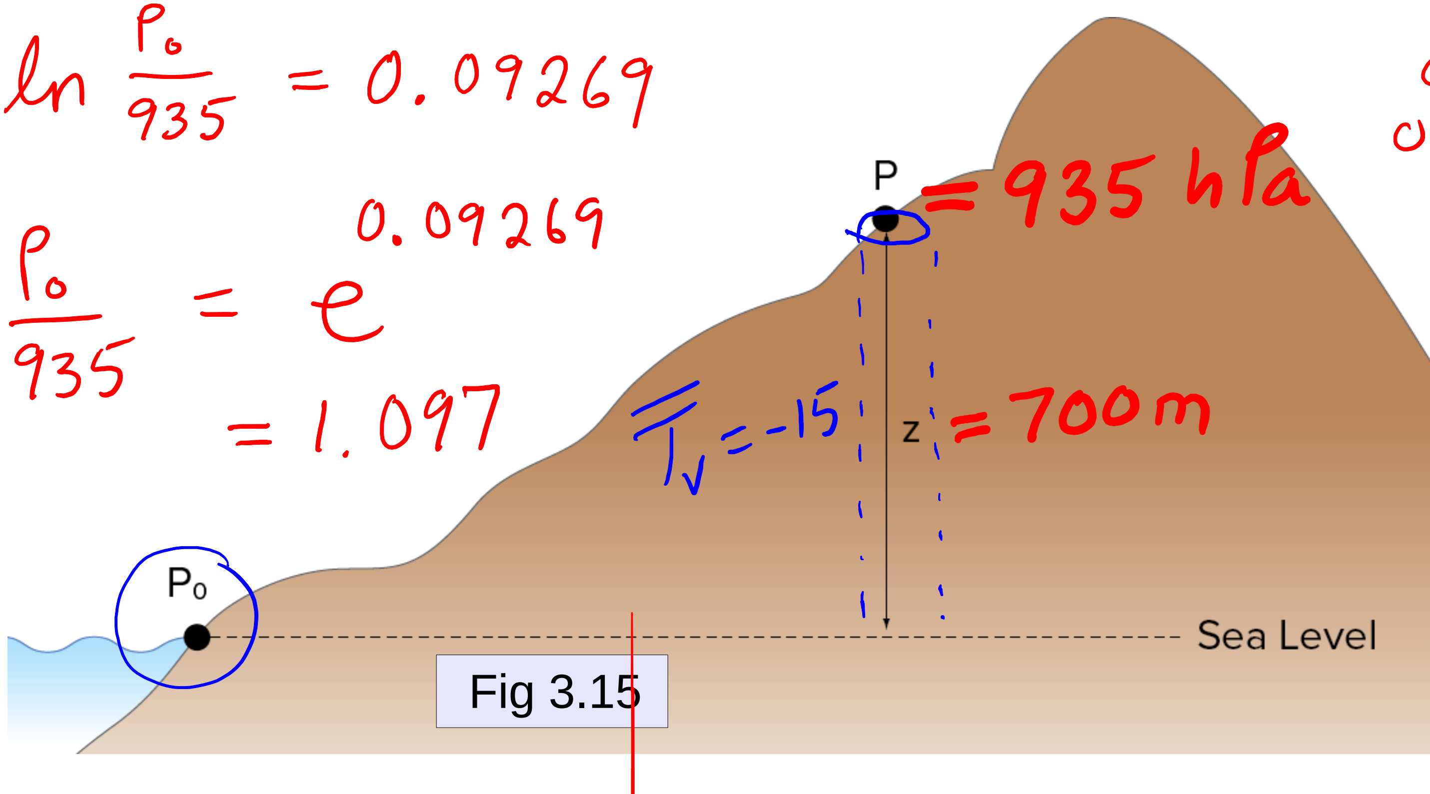


Fig 3.15

$$\Delta z \equiv z_2 - z_1 = \left[\frac{R_d \ \bar{T}_v}{g} \right] \ln \frac{P_1}{P_2}$$

- fix P_1=1000, P_2=500 hPa , then $\dfrac{R_d}{g} \ln \dfrac{P_1}{P_2} = 20.3$
- so $\Delta z \ [\mathrm{m}] = 20 \bar{T}_v$ or $\boxed{\Delta z \ [\mathrm{dam}] = 2 \bar{T}_v}$
- a 2 dam increase in layer mean thickness corresponds to a 1 K increase in mean temperature

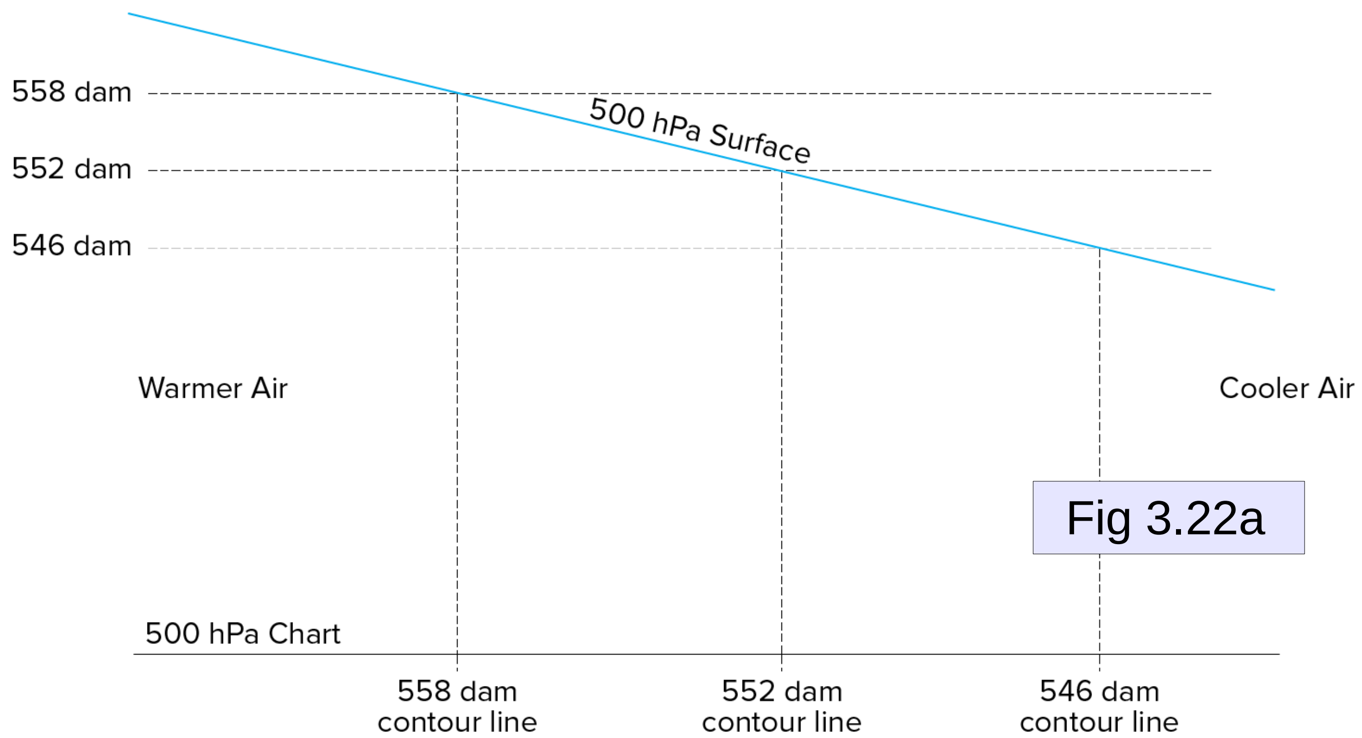


Fig 3.22a

- "thickness" from 1000 to 500 hPa, a surrogate for mean temperature of the lower troposphere
- 534-540 dam thickness band highlighted by fill
- Manitoba storm wrapped cold air southward through Saskatchewan

Why? It is considered that if the atmosphere were in a state to produce precipitation, then wherever $534 \le \Delta z \le 540$ dam freezing rain could occur

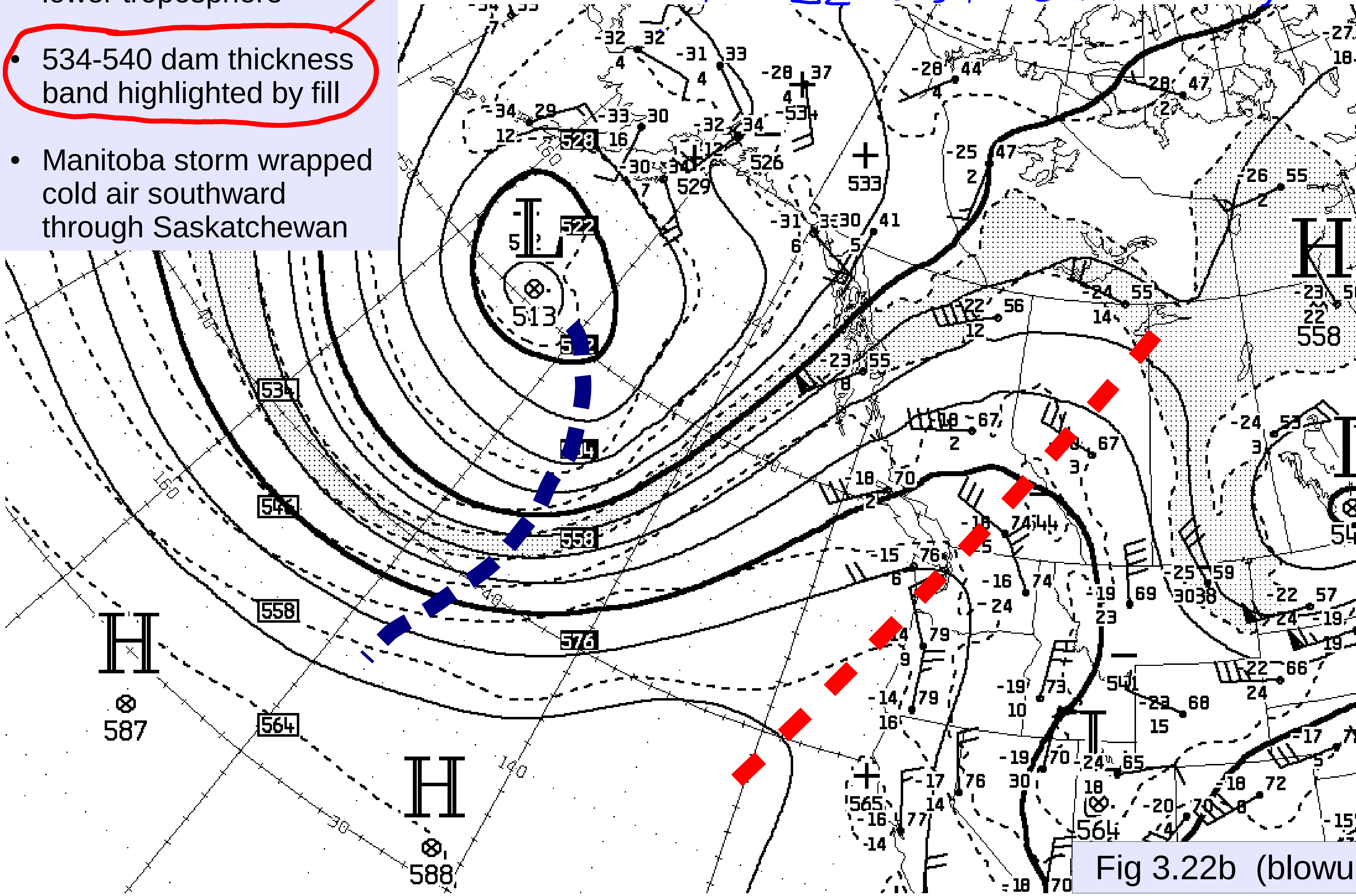
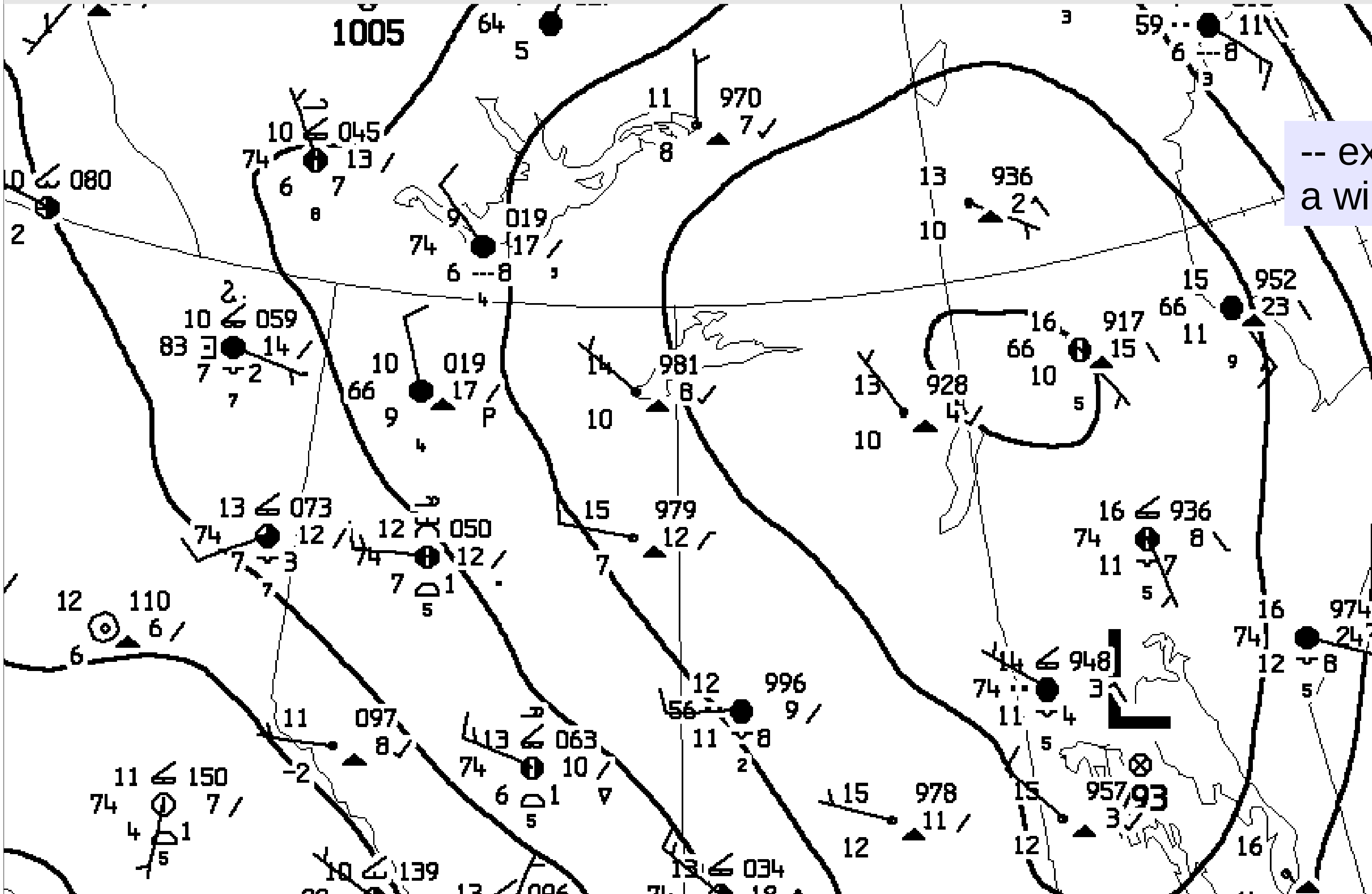


Fig 3.22b (blowup)

Close-packed isobars over Edmonton -- CMC surface analysis, 18Z Sunday 18 Sept. 2016



-- explain it's being a windy day

ie. Sunday's wind is very consistent with the isobar pattern over C. Alberta

Observed at: **Edmonton Blatchford**

Date: **2:00 PM MDT Sunday 18 September 2016**

Condition: **Not observed**

Pressure: **100.6 kPa**

Tendency: **rising**

Temperature: **14.4°C**

Dewpoint: **4.9°C**

Humidity: **53%**

Wind: **NW 35 km/h**

Lecture of 19 Sept.

- how to visualize an isobaric surface
- troughs, ridges, closed lows & highs – "topography" of the isobaric surface
- the hypsometric eqn
- using hypsometric eqn to define MSLP (sea-level corrected pressure)
- it is common on weather charts to display contours of 1000-500 hPa thickness as a surrogate for the mean temperature of the lower troposphere