ENV-320 Exam (Physics Exam Corrections)

Sample Solution P1:
The polar vortex forms over both poles in the respective winters (Austral winter in Antarctica) because very cold air masses cause a sinking motion and high pressure at the surface over the poles. At higher altitudes (such as depicted in Figure 1 and Figure 2) air is therefore advected towards the pole and the strong Coriolis force leads to the formation of circular movement and finally the vortex. The vortex will remain strong as long as the surface cooling (absence of light and therefore shortwave heating) persists.

Sample Solution P2a:
The polar vortex is stronger over Antarctica because the Antarctic land mass is approximately centered around the south pole and a much stronger surface cooling occurs over land than over sea. In addition the high elevation of the South Pole supports extremely low temperatures at the surface. This very cold surface at the pole is surrounded by (warmer) ocean water increasing the later temperature gradients. All these factors add up to an ideal situation for forming a strong vortex over Antarctica. At the north pole, less cooling over the Arctic ocean and stronger cooling over the land masses further away from the pole (Siberia and Canada) give a less ideal setting for vortex formation.
The winds have increased between May and June as can be seen from the intensified pressure gradient between the southern tip of South America and the Antarctic peninsula. The increased pressure gradient can be seen from the fact that the contours of geopotential in Figure 2 are coming closer together.

Sample Solution P2b:
Just re-type the equation and use the numbers given for f and Δn .There is about two contour line spacings between the southern tip of South America and the Antarctic peninsula in June und thus we have:

$$\frac{1}{0.00013} \cdot 2 \cdot \frac{5000}{2000000} = 38.5\ m/s$$

For May, it would be 19.2 m/s, assuming one contour line spacing.

Sample Solution P3a:

$$V = \frac{u^*}{\kappa} \ln\left({}^{z}/_{z_0}\right)$$

$u^* = \frac{V \cdot \kappa}{\ln\left({}^{z}/_{z_0}\right)}$ with $V = 17\ {}^{m}/_{s} \quad \kappa = 0.4 \quad z = 700m \quad z_0 = 0.2m$

$$u^* = 0.83328\ m/s$$

$$V_{15m} = \frac{0.83328}{0.4} \ln\left({}^{15}/_{0.2}\right) = 0.8994 m/s$$

Sample Solution P3b:
The surface exchange is proportional to the drag coefficient so we have:

$$e = 1 - \left(\frac{C_{0.005}}{C_{0.2}}\right) = 1 - \left(\frac{\ln\left(\frac{10}{0.2}\right)}{\ln\left(\frac{10}{0.005}\right)}\right) = 0.735$$

Sample Solution P4:
Once the parcel of air is forced up the mountain, the parcel will cool following the dry adiabatic lapse rate until the temperature of the parcel equals its dew point temperature. This occurs at the lifting condensation level (LCL). Above this point, if the parcel continues to rise (e.g., the mountain height exceeds the LCL), cloud droplets will form releasing latent heat from condensation and the air parcel will cool at a rate equal to the pseudo-adiabatic lapse rate that is lower in magnitude than the dry adiabatic rate. This lifting above the LCL leads to a loss of water vapor from the parcel. If the mountain range is high enough, the parcel may reach the level of free convection (LFC) where the temperature of the parcel equals the ambient temperature and the parcel is in unstable equilibrium, and depending on the perturbation, it can either continue to rise, will stay at approximately the same height or will start to descend. The usual case of descent after the mountain range can be caused by buoyancy (if the parcel is now heavier than the surrounding air) and/or by dynamical flow features such as a mountain wave. However, if we assume that the highest point of the mountain range is below this point, then once the parcel reaches this highest point, the parcel will start a descending motion on the downwind slope caused by the negative buoyancy (pointing downwards) because the temperature of the parcel is lower than the temperature of the surrounding air. The temperature of the parcel during this descending motion should follow the dry adiabatic lapse rate, reaching eventually a new point of equilibrium in which the parcels temperature equals the ambient temperature. The loss of water vapor will lead to a warmer and drier air parcel with respect to its original temperature and humidity.