

The Tephigram

Introduction

Meteorology is the study of the physical state of the atmosphere. The atmosphere is a heat engine transporting energy from the warm ground to cooler locations, both vertically and horizontally. The driving force is solar radiation. Shortwave radiation is absorbed primarily at the surface; the working fluid is the atmosphere, which distributes heat by motion systems on all time and space scales; the heat sink is space, to which longwave radiation escapes.

The Tephigram is one of a number of thermodynamic diagrams designed to aid in the interpretation of the temperature and humidity structure of the atmosphere and used widely throughout the world meteorological community. It has the property that equal areas on the diagram represent equal amounts of energy; this enables the calculation of a wide range of atmospheric processes to be carried out graphically. A blank tephigram is shown in figure 1; there are five principal quantities indicated by constant value lines: pressure, temperature, potential temperature (θ), saturation mixing ratio, and equivalent potential temperature (θ_e) for saturated air.

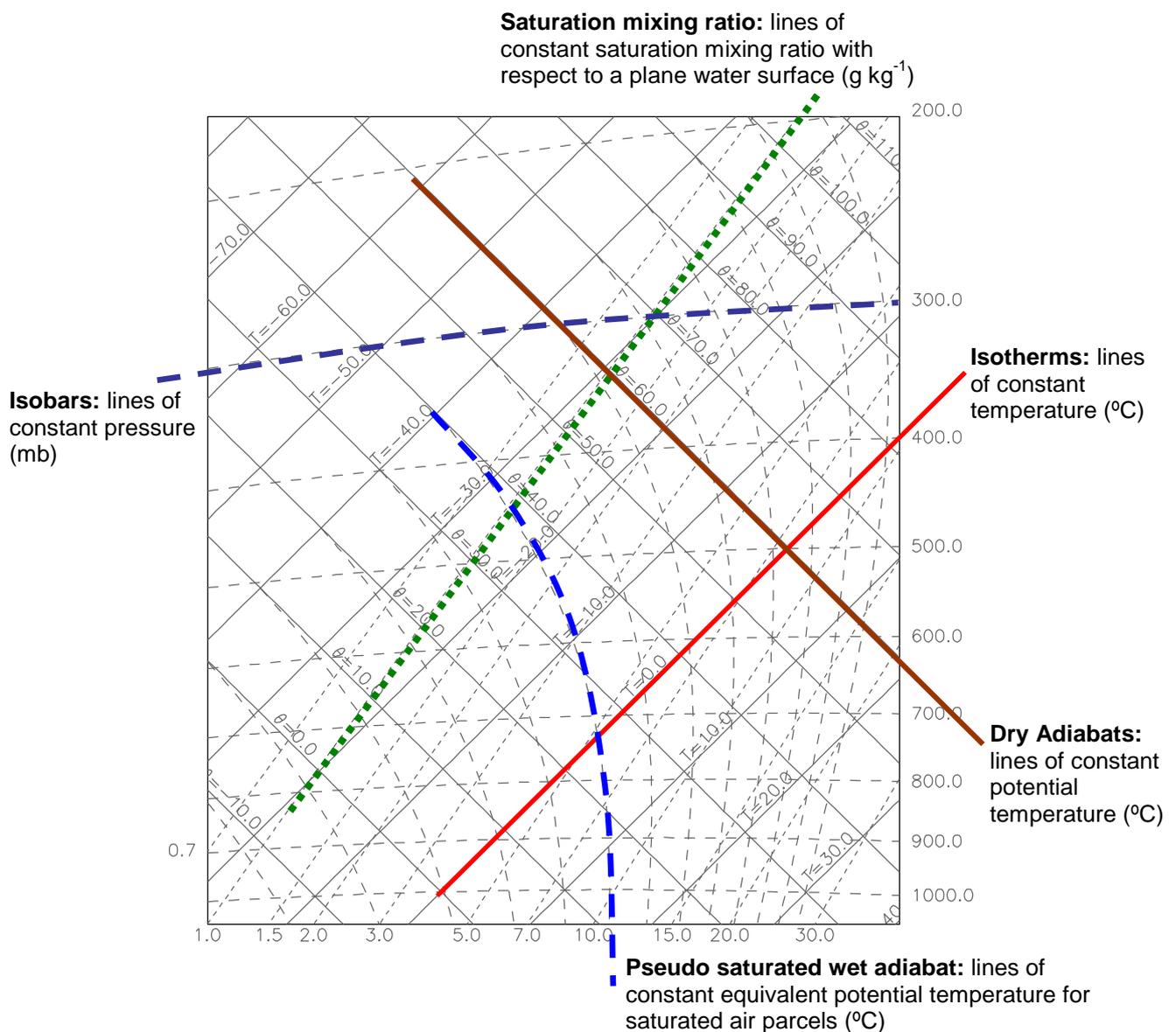


Figure 1. The tephigram, with the principal quantities indicated.

The principal axes of a tephigram are temperature and potential temperature; these are straight and perpendicular to each other, but rotated through about 45° anticlockwise so that lines of constant temperature run from bottom left to top right on the diagram. This rotation makes lines of constant pressure almost horizontal, though gently curving down towards bottom left, so that altitude increases from bottom to top of the diagram. The pseudo-saturated wet adiabats (often simply called saturated adiabats) are the most visibly curved lines on the diagram, being almost vertical at high pressures (near the surface, or bottom of the diagram) and approaching the dry adiabats with decreasing temperature. Saturated air cools at a lesser rate than dry air due to the release of latent heat as water condenses out; and the rate varies due to the non-linear variation of saturation vapour pressure with temperature. The saturated adiabats usually extend only to the -40°C level; at temperatures this low the air can hold so little water vapour that it is essentially dry; at lower temperatures the saturated adiabat is assumed equal to the dry adiabat. Finally, lines of constant saturation mixing ratio run from bottom left to top right.

Note that while the temperature and both dry and saturated adiabats usually have values indicated in degrees celcius, the potential and equivalent potential temperature should strictly be Kelvin, and all calculations involving them should use values in Kelvin.

The data plotted on a tephigram are the air temperature (T) and dew-point temperature (T_d) plotted against pressure. These are usually obtained from the measurements made by a radiosonde.

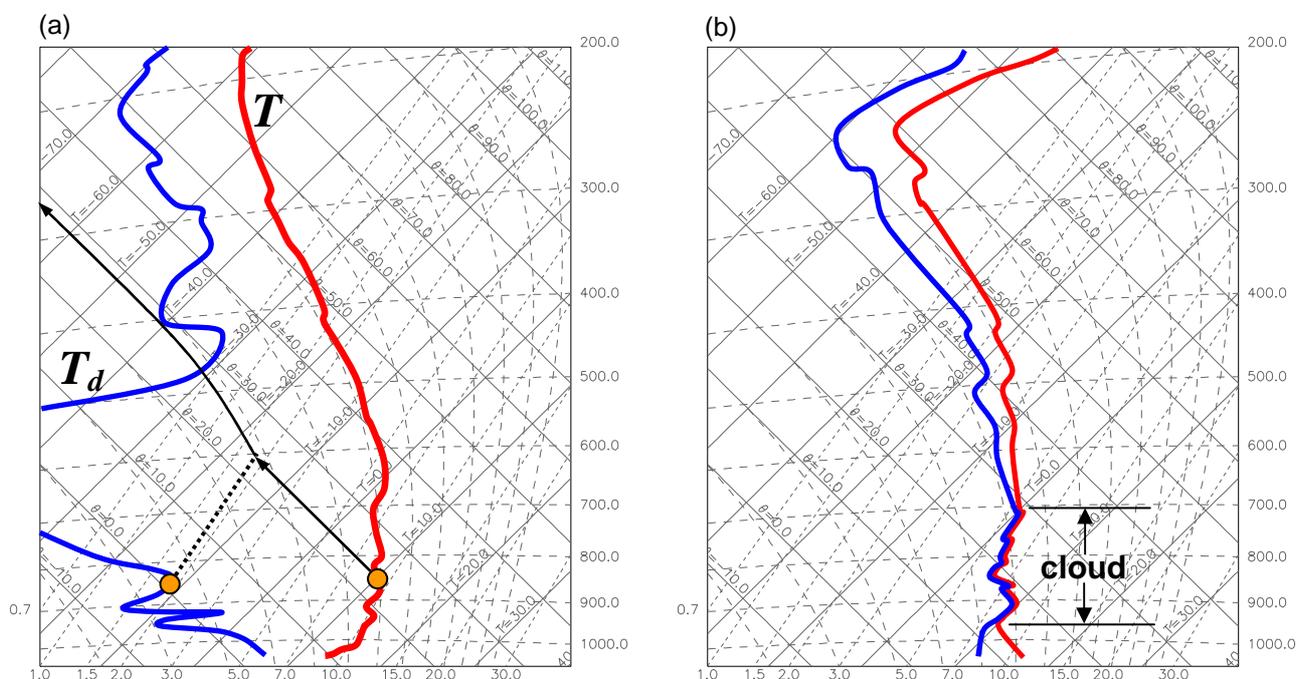


Figure 2. Example tephigrams

The thermodynamic properties of a parcel of air are defined by corresponding points on the temperature and dew point curves *at the same pressure*. The dew point is the temperature at which water vapour would first condense out if a parcel of air were cooled at constant pressure; this defines the actual mixing ratio of the parcel of air (about 1.9 g kg⁻¹ for the example in figure 2(a)). If the parcel of air were to be cooled by adiabatic lifting – for example in flow forced to rise over hills – it would trace out a path on the tephigram along a dry adiabat until it reached the point where the saturation mixing ratio fell to the actual mixing ratio of the parcel; at this point water would condense out to form cloud; this is known as the *lifting condensation level*. If the parcel were lifted higher still, its temperature would follow the saturated adiabat as shown in figure 2(a). Regions where the temperature and dew point temperature are equal indicate where the radiosonde passed through saturated air – cloud (Figure 2(b)).

The stability of the atmosphere can be determined from the temperature profile on a tephigram, along with the approximate vertical extent of any convective lifting and of consequent cloud formation. Locally, the stability of a layer of air depends upon its density with respect to the air

around, above, and below it; and the extent of changes in density resulting from vertical displacements. A parcel of air that is denser than the air around it will sink downwards, a parcel that is less dense than the air around it will rise upwards. A parcel is said to be stable if, following a forced vertical displacement its density with respect to the ambient air at the level it has moved to is such that it would tend to sink or rise back towards the level at which it started. It is said to be unstable if after a small vertical displacement its density with respect to ambient was such that it would move away from its original position. Changes in air density depend on changes in temperature and water vapour mixing ratio, both of which are accounted for by following temperature adiabatic temperature paths on the tephigram. A parcel of air that is warmer than the air around it is less dense and therefore buoyant, and the layer said to be unstable; air that is cooler than the air around it is denser and therefore negatively buoyant, and the layer is said to be stable. The idealised examples below illustrate several different classes of stability.

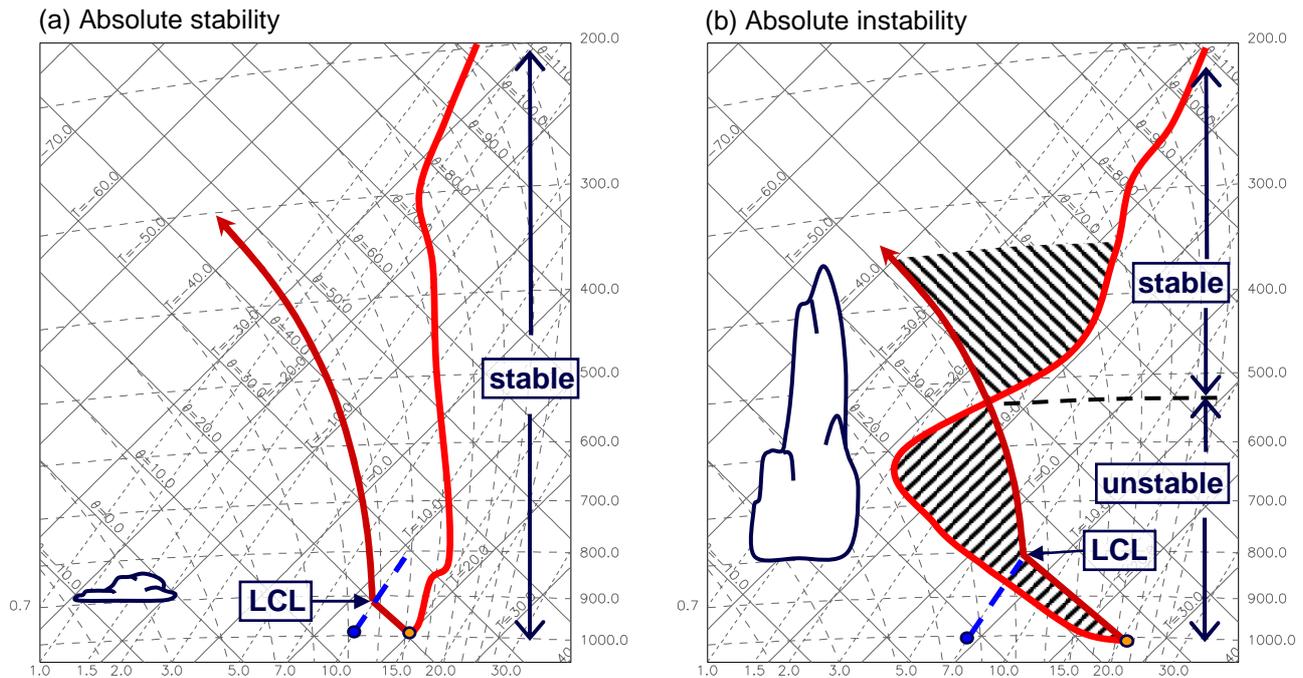


Figure 3. Idealised examples of absolute static stability (a) and instability (b) of air at the surface

Figure 3(a) shows a stable temperature profile: if air at the surface (or in this case from any level) is forced to rise, it cools dry adiabatically until it reaches its lifting condensation level, and along a saturated adiabat with any further lifting. At all points the parcel is cooler, and hence denser, than the ambient air at the pressure level to which it has been lifted. It is thus stable, tending to sink back towards the level at which it started. Lifting here must be externally forced. If air from above the surface were to be forced downwards, it would become warmer and hence less dense than the air around it, and tend to rise back upwards. Under these conditions a parcel of displaced air, moving back towards its level of origin may overshoot – it may oscillate about its original level a number of times, slowly losing energy. Such oscillations are called gravity waves, and often occur in where stable air is forced up over orography. If cloud forms in the lifting part of the wave, the waves can be clearly seen in satellite imagery.

Figure 3(b) shows a case for which the near-surface air is unstable. Adiabatic lifting results in an air parcel warmer, thus less dense, than the surrounding air; this will be positively buoyant and will rise upwards as a convective plume. It rises dry adiabatically until its lifting condensation level, then along a saturated adiabat. In the case shown a strong temperature inversion exists at the top of the boundary layer at about 600 mb; within the inversion the ambient temperature increases with altitude. When the rising plume of air reaches the point at which its temperature equals that of the surrounding air it is no longer positively buoyant, this is a level of neutral buoyancy. The inertia of the plume of air will, however, cause it to overshoot the level of neutral buoyancy; it will then become cooler than the surround air and negatively buoyant. The maximum extent of the overshoot is reached when all the energy gained within the region of positive buoyancy has been used up in

ascending through the region of negative buoyancy – this is achieved when the two hatched areas between the curves are equal. In the real world this maximum overshoot is rarely reached since energy is lost by friction and in forcing upwards through the overlying air at all points during the ascent; but it remains a useful guide to the approximate maximum height of updraughts.

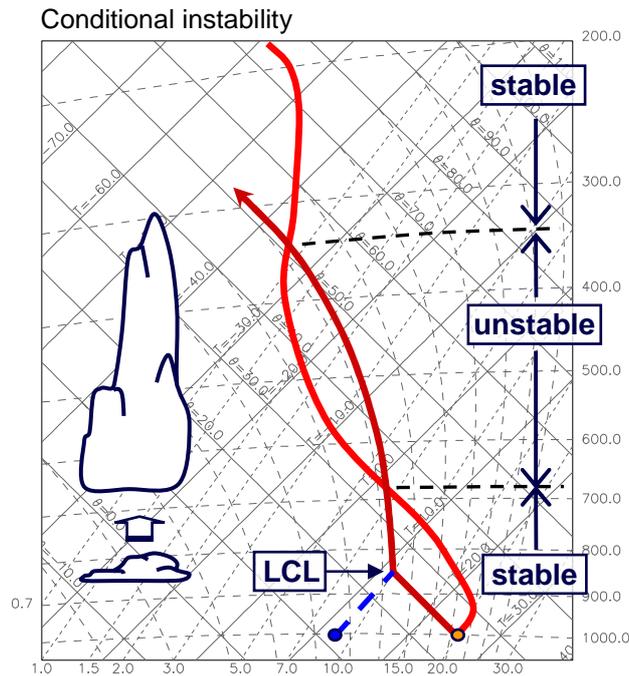


Figure 4. Idealised example of conditional instability

Figure 4 shows an example of conditional instability. A parcel of air lifted adiabatically from the surface remains cooler than its surroundings, and thus stable. If the parcel remained dry, it would remain stable no matter how high it was lifted; however, in this case the LCL is reached at around 850mb, and the rate of cooling decreases to that of the saturated adiabat. The saturated adiabat intersects the environmental temperature profile at about 680 mb; if the parcel of surface air is forced to lift past this point it becomes warmer than its surroundings, and thus positively buoyant, and will continue to lift convectively until about the 340mb level, above which it becomes negatively buoyant again. This case is described as conditional instability, because reaching the point of instability is conditional on forced lifting through a region on stability.