2. Meteorology

2.1 Meteorological Principles

The hydrological response of the Mawddach catchment is clearly dependent on the timing, distribution and intensity of rainfall during individual storm events, and the conditions of the ground as a result of antecedent rainfall. Aspects of meteorology affecting rainfall over western Britain are outlined in this section. An important component of the Mawddach hydrology project is the development of a rainfall model for the catchment. The setting-up and evaluation of this model will be discussed in section 2.4.

Rainfall in western Britain is mainly associated with the passage of low pressure centres termed *Mid-latitude Cyclones* or *Depressions*. To understand the air flow patterns around a low pressure centre, it is necessary to consider the effects of the Coriolis force (fig.2.2). This is an apparent force due to the rotation of the earth which deflects moving bodies towards the right in the northern hemisphere. Air will tend to move from areas of high pressure towards areas of low pressure. As air is flowing towards a low pressure centre in the northern hemisphere, it will experience Coriolis deflection to the right. This induces a spiral track towards the centre (fig.2.1a). If the pressure forces should become exactly balanced by the Coriolis force, winds will circulate in a direction perpendicular to the pressure gradient and are termed *geostrophic winds* (fig.2.1b). In the southern hemisphere the direction of rotation towards a low pressure centre is reversed to produce clockwise motion.
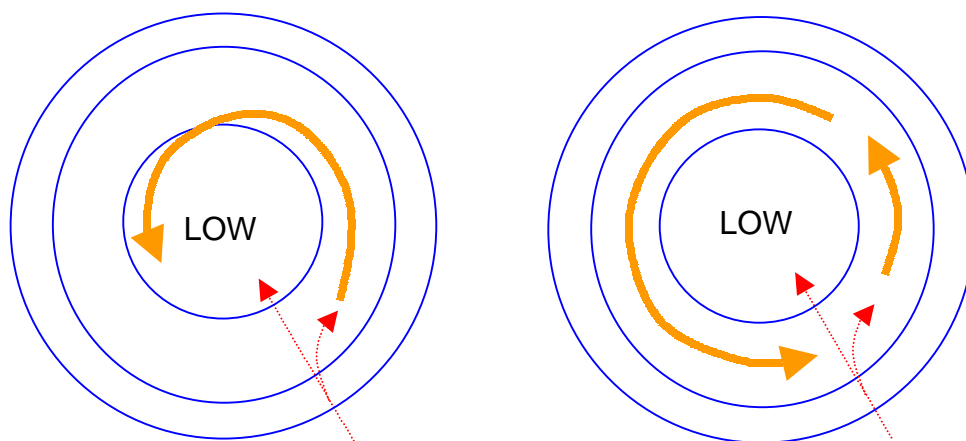
Figure 2.1a (left): Coriolis deflection of winds approaching a low pressure centre in the northern hemisphere.
Figure 2.1b (right): Geostrophic winds developed around a low pressure centre.

The Coriolis force is an apparent force experienced by a body due to the rotation of the Earth.

Imagine a body moving due south relative to a fixed frame of reference in space (a).
As the Earth rotates, the track of the body over the Earth's surface maps out a curve (b-c).

(a)

(b)

(c)

(d)
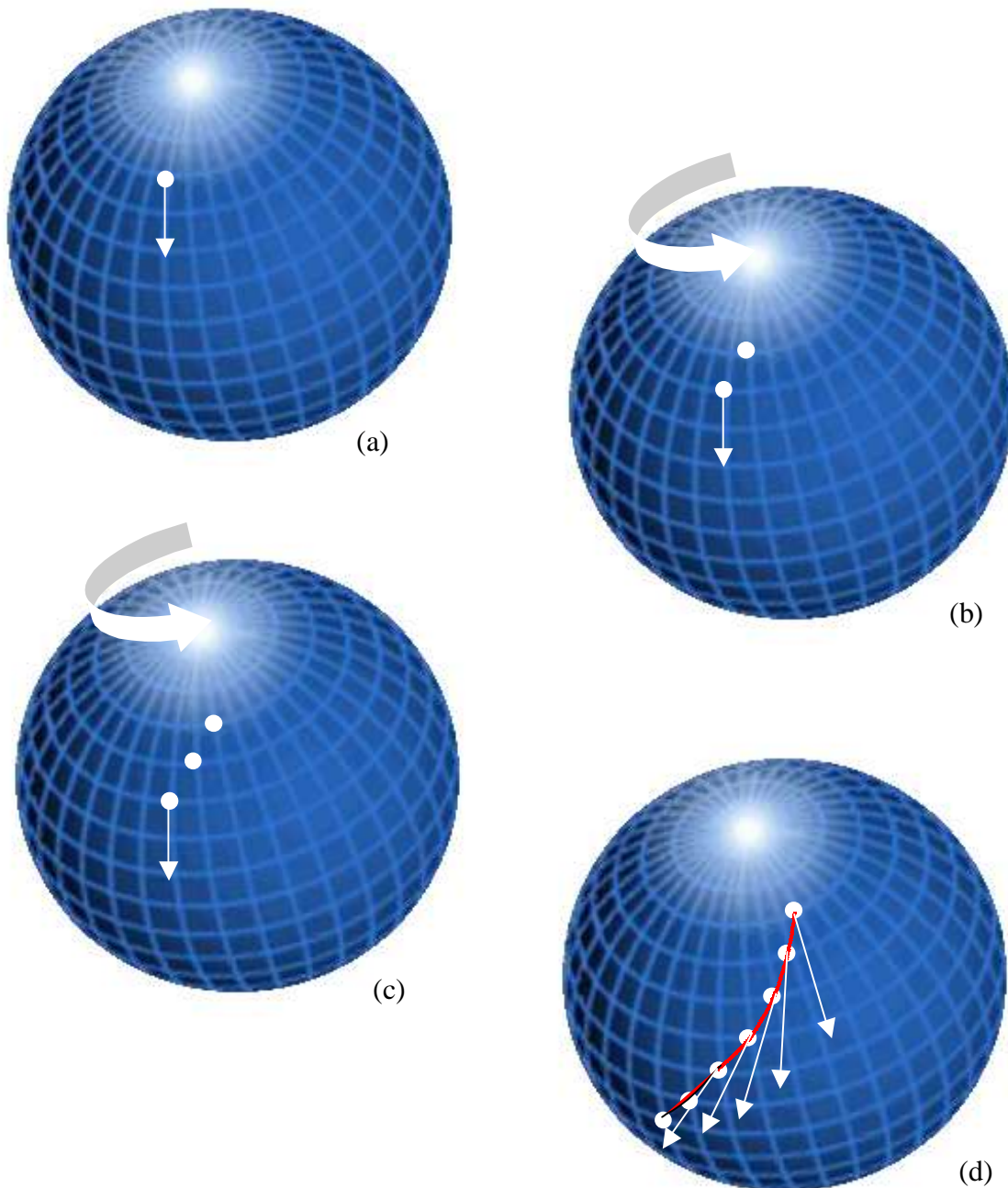
It would appear that a force was operating which was continuously accelerating the body to the right of its track (d) in the northern hemisphere, or to the left in the southern hemisphere. This apparent force is termed the **Coriolis force**.

Figure 2.2: The Coriolis effect

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The satellite image of fig.2.3 shows typical cyclonic circulation about a low pressure centre in the mid-Atlantic.
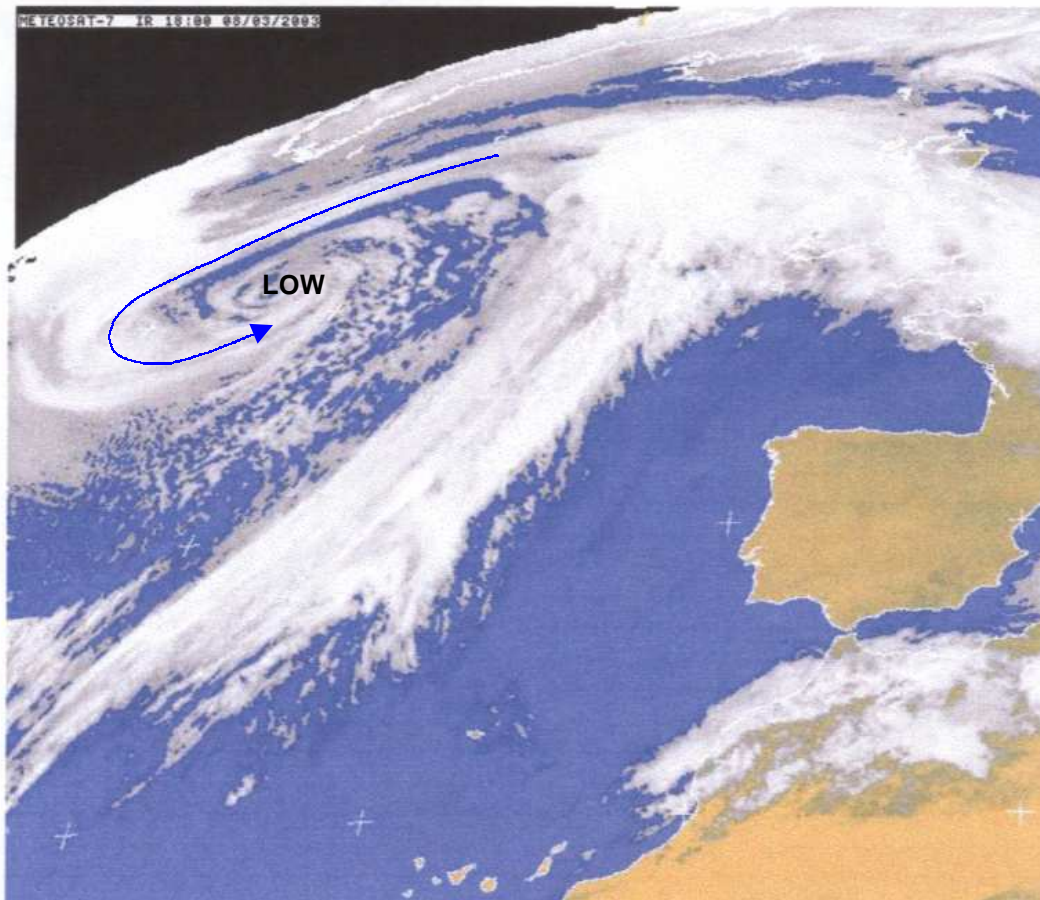

Figure 2.3: Meteosat-7 image for 18:00h, 8 March 2003.
Dundee Satellite Receiving Station

Weather processes take place within the lowest layer of the atmosphere, the *troposphere*, which varies in thickness between 8 and 17km. The circulatory system of the earth controls weather patterns; the driving force is the tendency to transfer heat from equatorial areas towards cold polar areas by the movement of air masses. A model for the large scale air motions of the planet was first proposed by Hadley in 1735, and has been progressively refined as more detailed observational data became available. A vertical section through the atmosphere of the northern hemisphere is given in fig.2.4. Within the tropical belt, large convective cells carry warm air northwards at high level, with a return flow near the surface.
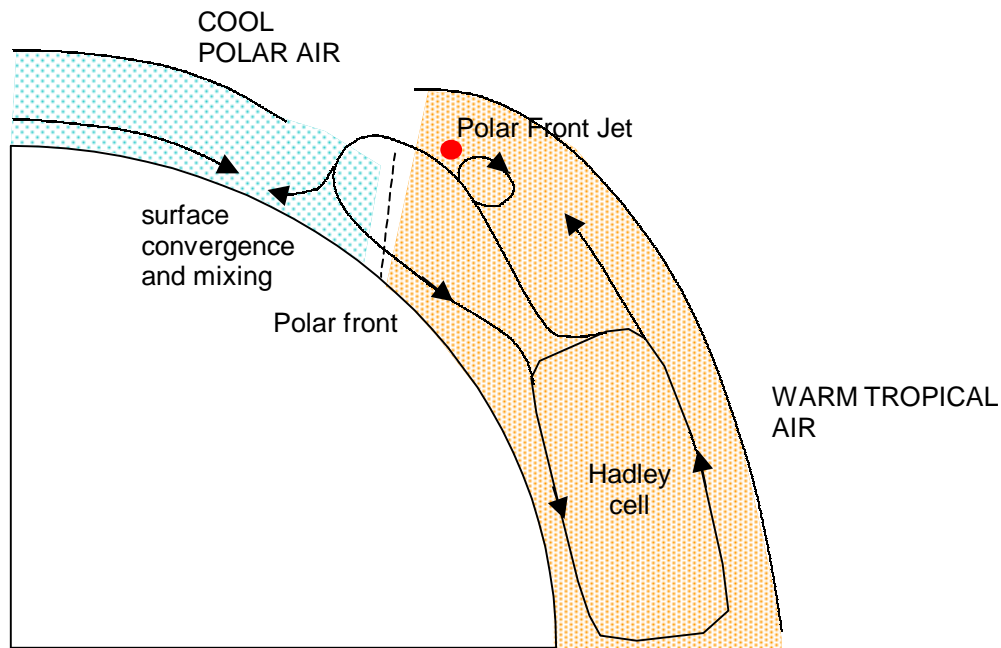
Figure 2.4: Flow of air masses in a generalised vertical section of the northern hemisphere. After Barry and Chorley 1976.

Air movements in mid-latitudes are more complex. The cold Polar air mass is separated from the warm Tropical air mass along a narrow zone called the Polar Front. Since warm air expands upwards in comparison to cold air, the troposphere is thicker for the Tropical air mass and thinner for the Polar air mass. The Polar Front is a *baroclinic* zone across which the air pressure changes rapidly. This is a result of the different thicknesses of the Polar and Tropical air columns which extend above any chosen reference level in the troposphere.

The Polar Front Jet

A high velocity channel known as the Polar Front Jet exists a short distance to the south of the Polar Front and near to the top of the troposphere. This jetstream flows from west to east across the Atlantic, with velocities reaching 250km/h. The mean winter position of the Polar Front Jet is shown in fig.2.5.
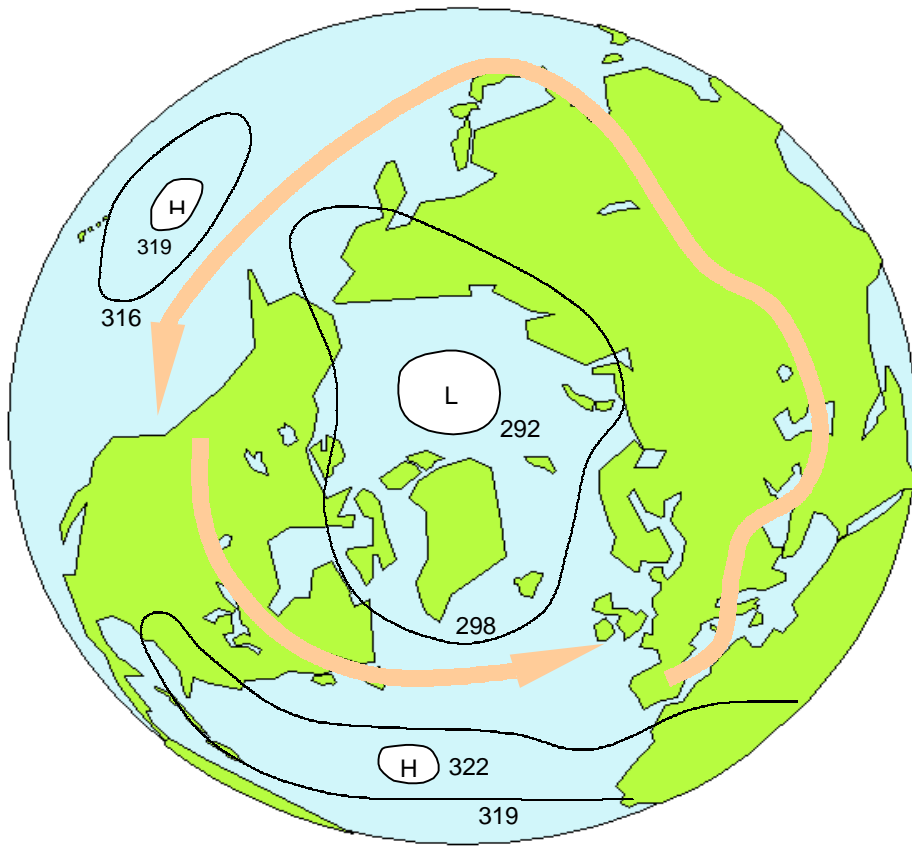
Figure 2.5:
Average pressure distribution and location of the Polar jet stream: winter. Heights of 700mb surface (geopotential hectometres), after Barry and Chorley, 1976

319
316
H 319
L 292
298
H 322
319

cyclonic vorticity and removal of air aloft

Tropical tropopause

TROPICAL AIR

Polar tropopause

Subtropical jetstream

Polar front jetstream

ascent

POLAR AIR

LOW PRESSURE

warm front

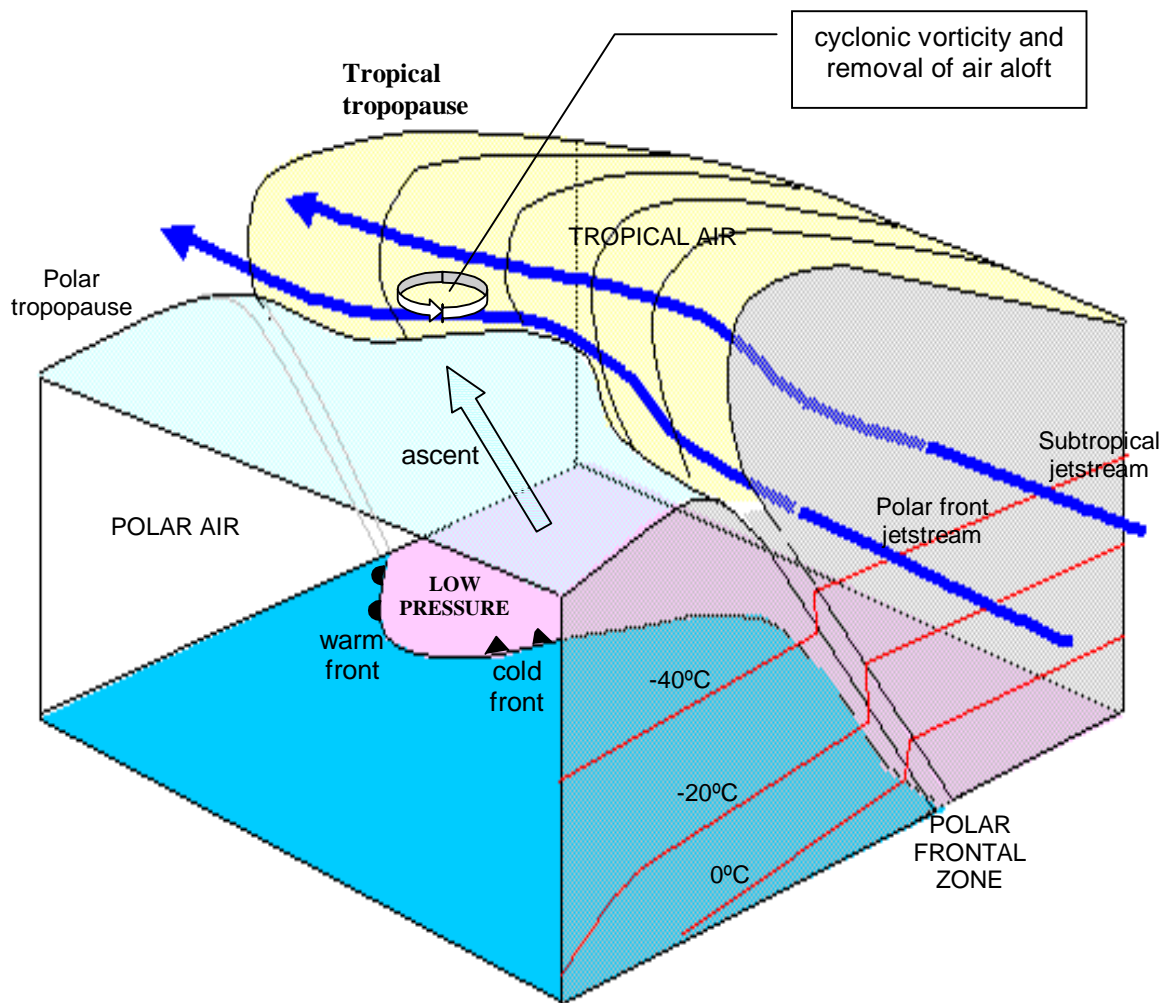cold front

-40°C

-20°C

0°C

POLAR FRONTAL ZONE

Figure 2.6: Diagramatic cross section through the Polar frontal zone of the North Atlantic

The Polar font jet stream provides a mechanism for dissipation of angular momentum, and appears to play a central role in controlling weather patterns of the North Atlantic.

Rotating bodies must obey the principle of conservation of angular momentum. For air masses, the total vorticity about a vertical axis is made up of the sum of the Coriolis force f related to the earth's spin, and vorticity ζ about a local point such as a high or low pressure centre. Overall, angular momentum is conserved. This is expressed mathematically as:

$$\frac{d(f + \varsigma)}{dt} = 0$$

If air moves poleward so that f increases, then ζ tends to decrease correspondingly, producing anticyclonic vorticity and inducing curvature of the air stream to the right. If air moves towards the equator so that f decreases, then ζ correspondingly increases, producing cyclonic vorticity and inducing curvature of the air stream to the left. This process gives rise to the large waves seen in the north Atlantic jetstream, known as Rossby waves
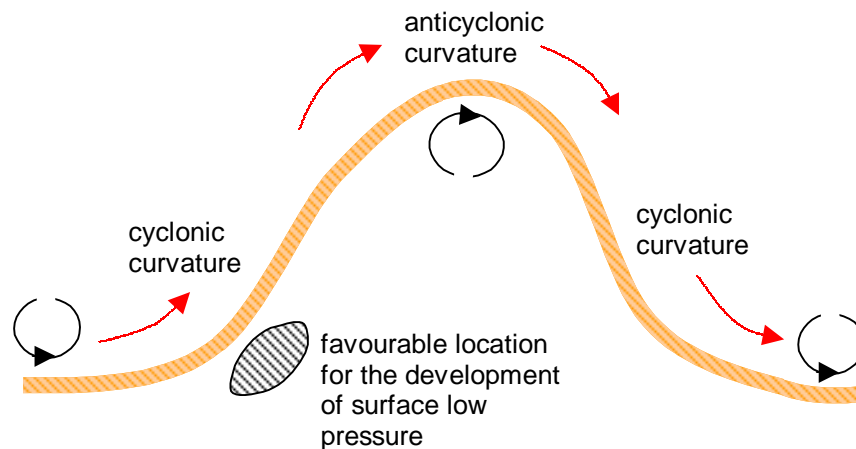

Figure 2.7: Induced vorticity along Rossby long waves.

Fig.2.6 illustrates the relationship between the jet stream pattern in the upper troposphere and the surface distribution of weather fronts. A *front* is a boundary between air masses with different physical properties. The Polar air mass is colder and generally has lower pressure. The Tropical air mass is warmer and generally has higher pressure.

A key factor in production of weather systems is the ability of cyclonic circulations in the upper troposphere to disperse air outwards. The air beneath cyclonic circulations tend to rise to compensate, forming low pressure centres at the earth's surface called *depressions.* Rising air provides the conditions necessary for condensation of atmospheric moisture and rain production. A particularly favourable location for the development of cyclonic circulations is just ahead of the Rossby wave where it penetrates furthest to the south. Anticyclonic circulations, by contrast, draw air inwards at high level. Beneath an anticyclonic circulation, the air column tends to sink, producing a high pressure centre at the earth's surface and stable, dry conditions.

The jet stream is often broken into a series of wider segments of high velocity, known as *jet streaks*, separated by narrower zones of lower velocity. It is known that low pressure centres can develop where air flows enter a jet streak from the equatorial direction (right entry sector), and where air flows leave a jet streak in the polewards direction (left exit sector). Examples are shown in fig.2.8. At this time on 11 November 2005, there were indeed low pressure frontal systems located to the west of Ireland and over Spain. These lows would be favourable for the ascent of air and production of rainfall.
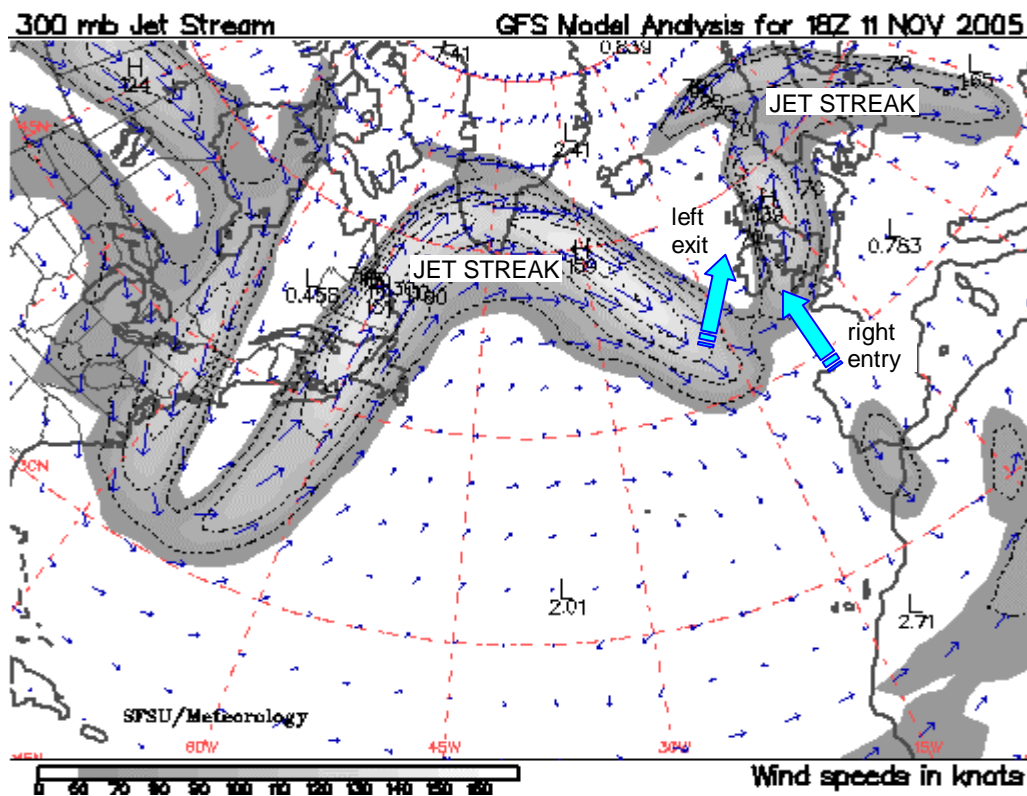

Figure 2.8: Jet stream map, identifying favourable locations for low pressure development in relation to jet streaks

The Polar jet stream is not fixed in position over the Atlantic region. A cycle of changes in the jet stream is commonly observed:

1. Commencement with a sub-linear pattern, known as a *zonal* jetstream
2. Sigmoidal waves are initiated
3. The sigmoidal waves develop with greater amplitude to form a *meridional* jetstream pattern
4. The jet stream continuity finally breaks down into a pattern of circular vortices.

The cycle will then be repeated with the formation of a new sub-linear pattern. Examples of zonal and meridional jetstream patterns during the period 24-26 October 2005 are shown in fig.2.9.
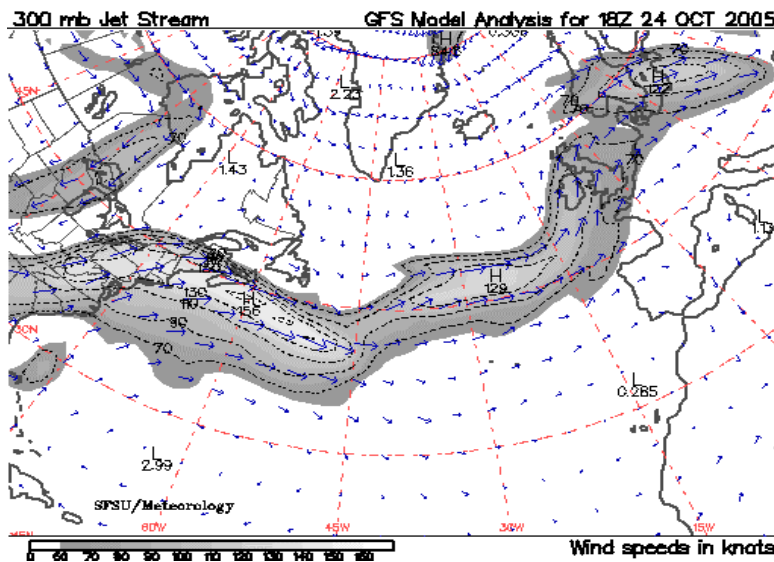

Figure 2.9(a).
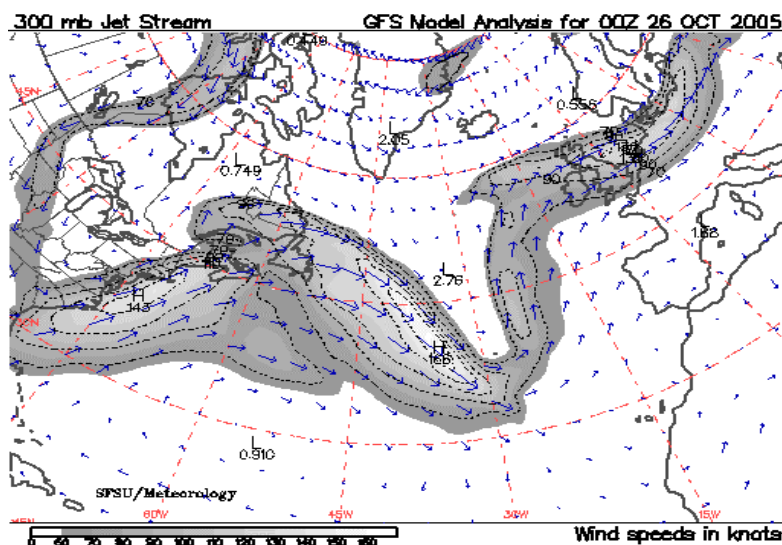**Zonal Polar Front Jet pattern in the Mid Atlantic,**
24 October 2005

Figure 2.9(b).
Meridional Polar Front Jet pattern in the Mid Atlantic,
26 October 2005

source: California Regional Weather Server

Theories of midlatitude cyclone structure

Early work in determining the structure and evolution of low pressure centres along the Polar Front was carried out by Bjerknes and Soldberg (1922). The theory, known as the *Norwegian Model*, envisages the development of a wave in the Polar Front which develops into distinct warm and cold fronts enclosing a wedge of tropical air termed the *warm sector* (fig.2.10).
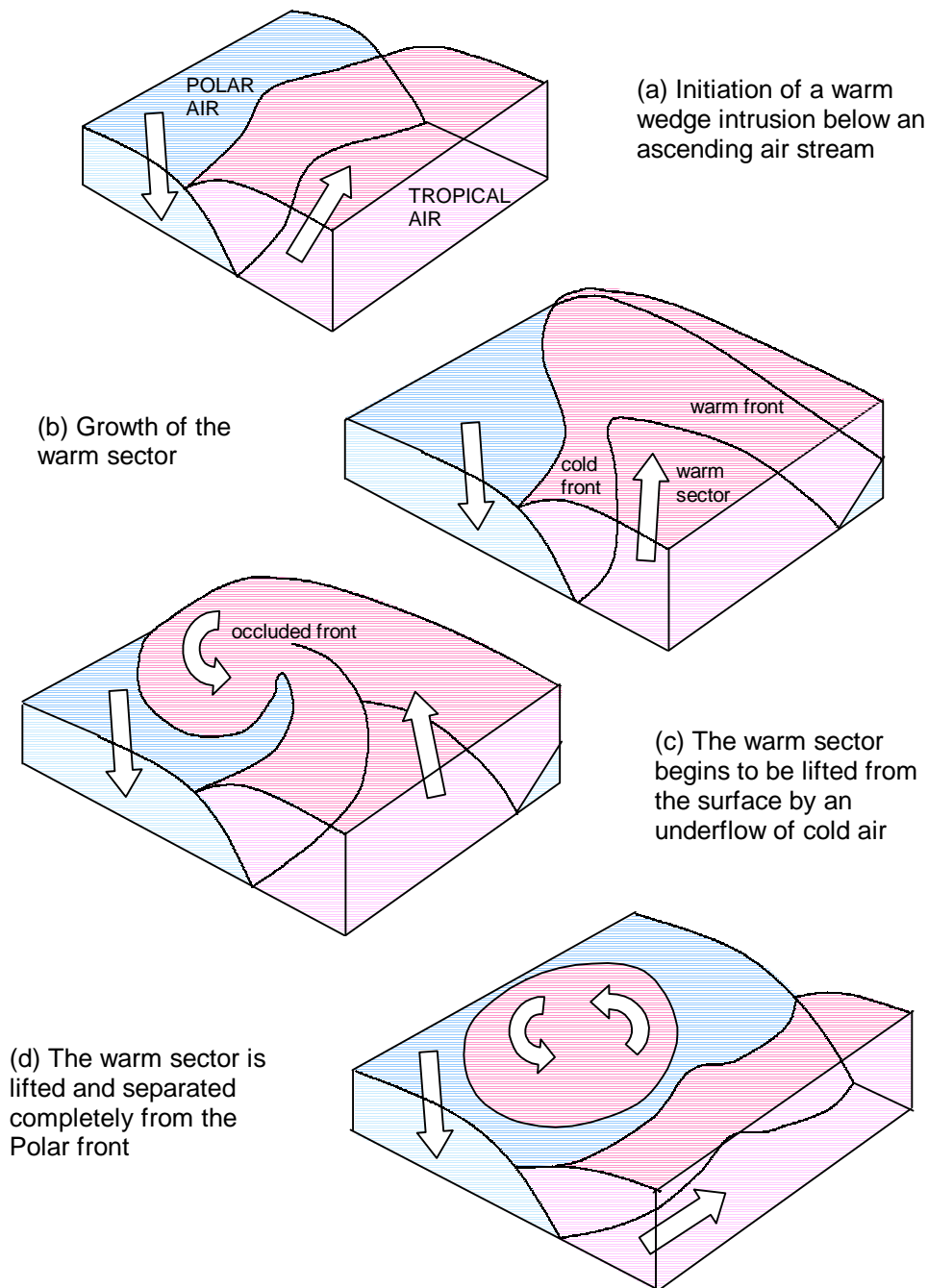
(a) Initiation of a warm wedge intrusion below an ascending air stream

POLAR AIR

TROPICAL AIR

(b) Growth of the warm sector

warm front

cold front

warm sector

occluded front

(c) The warm sector begins to be lifted from the surface by an underflow of cold air

(d) The warm sector is lifted and separated completely from the Polar front

Figure 2.10: Evolution of the frontal system of a mid-latitude depression, after Barry and Chorley, 1976

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A feature of the Norwegian model is that the cold front to the rear of the warm sector advances more rapidly than the warm front, and gradually overtakes the warm sector. The warm sector is progressively closed by occlusion of the cyclonic system, with warm air lifted from the ground and separated from the Polar Frontal zone.

With improved satellite imagery, it became increasingly evident that features of many cyclonic centres developed over the North Atlantic do not fit well with the Norwegian Model.

Browning and Hill (1984) identify three cases where cyclonic vortices develop in association with a Polar air mass low pressure centre:

- In the first case, the low pressure centre lies close to the Polar Front. In this situation, warm and cold fronts develop in a pattern consistent with the Norwegian model as shown in the example of fig.2.11.
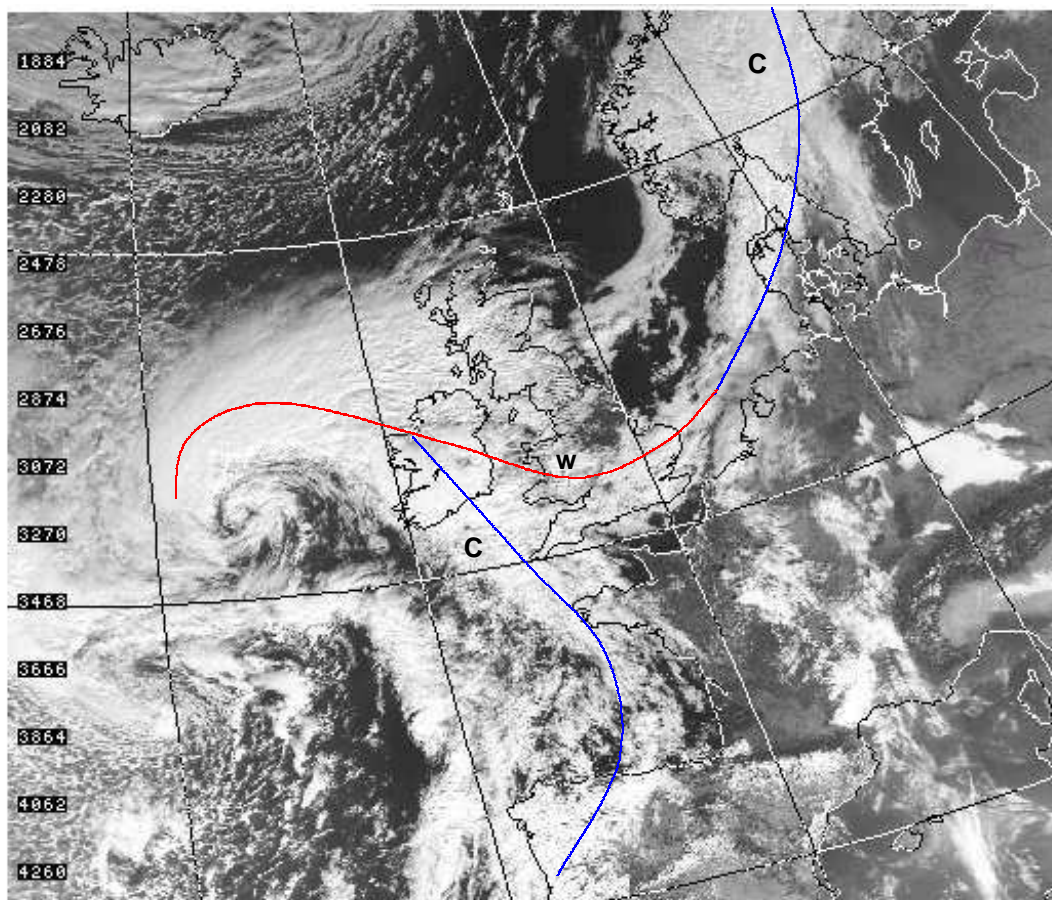

Figure 2.11: Satellite photograph for approximately mid-day on 29 October 2005, showing the development of warm-sector depression at the Polar Front (c: cold front, w: warm front). Dundee Satellite Receiving Station

- A second case is the situation where the low pressure centre lies within the Polar air mass well away from the Polar Front. No interaction occurs with the Polar Front, and an isolated vortex develops with a 'comma cloud' pattern as in the example of fig.2.12:
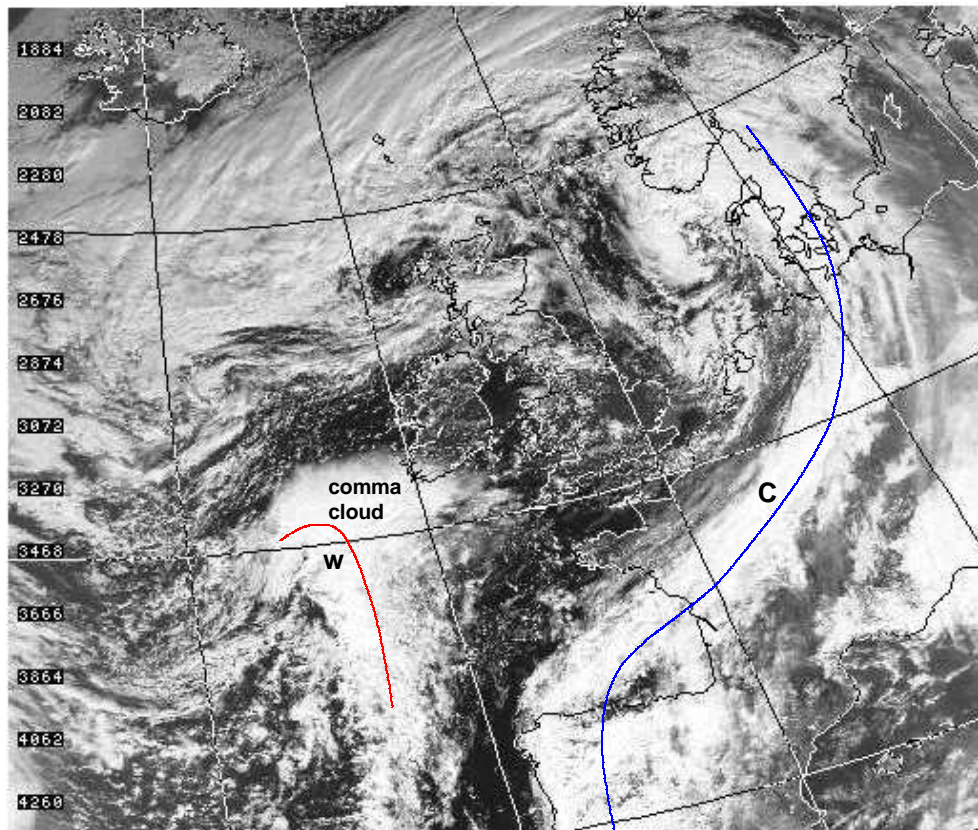

Figure 2.12: Satellite photograph for approximately mid-day on 20 October 2005 showing comma cloud development around a Polar air mass cyclonic centre. (c: cold front, w: warm front). Dundee Satellite Receiving Station

Significant to the model of Browning and Hill is the concept of the Polar air mass low pressure centre being the driving force for cyclonic convergence, uplift and air removal into the upper troposphere. Air flow within the vortex follows an ascending spiral path. Whether or not warm air from the Polar Front becomes entrained in the cyclone will depend on the proximity of initial vortex development to the surface position of the Polar Front. This mechanism is in contrast to the perceived driving force for the classical Norwegian model of cyclonogenesis. The Norwegian model assumes that the initiation and growth of wave disturbances along the Polar Front produces a warm sector intrusion, by analogy with the initiation and growth of meanders in a river.

- A third case represents an intermediate position for the low pressure centre. An initially isolated vortex draws in warm air from the Polar Front at a late stage in its development, as shown in fig.2.13.
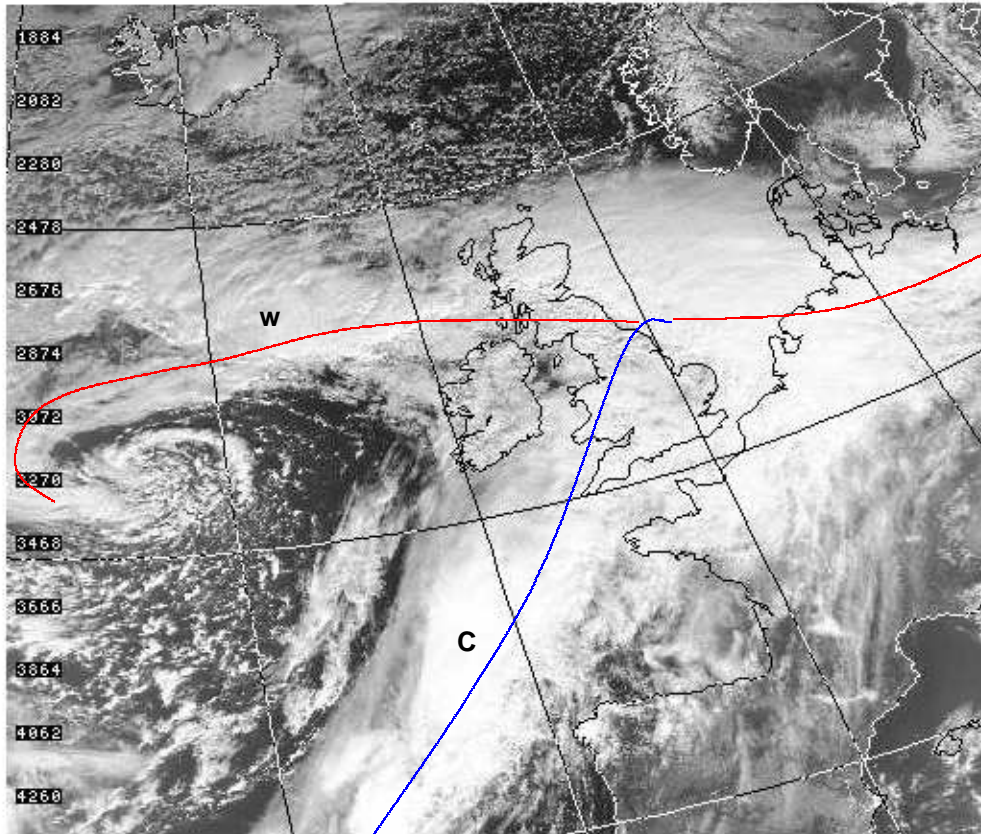

Figure 2.13: Satellite photograph for approximately mid-day on 24 October 2005, showing late stage linking of a cyclonic centre to the Polar Front. (c: cold front, w: warm front). Dundee Satellite Receiving Station

A further model for frontal system development has been developed by Shapiro and Keyser (1990). This is summarised in an illustration from their paper, reproduced here as fig.2.14. Early development of a Polar Frontal cyclone in the Shapiro-Keyser model resembles that of the Norwegian model, with formation of warm and cold fronts along a wave structure in response to a developing vortex in the Polar air mass. The cold front soon becomes detached from the warm front and remains roughly perpendicular to the warm front for much of the development sequence. This limits the closure of the warm sector. The warm front is considered to extend beyond the warm sector to form a *'T-bone'* structure, and bends back on itself towards the centre of the low pressure vortex.
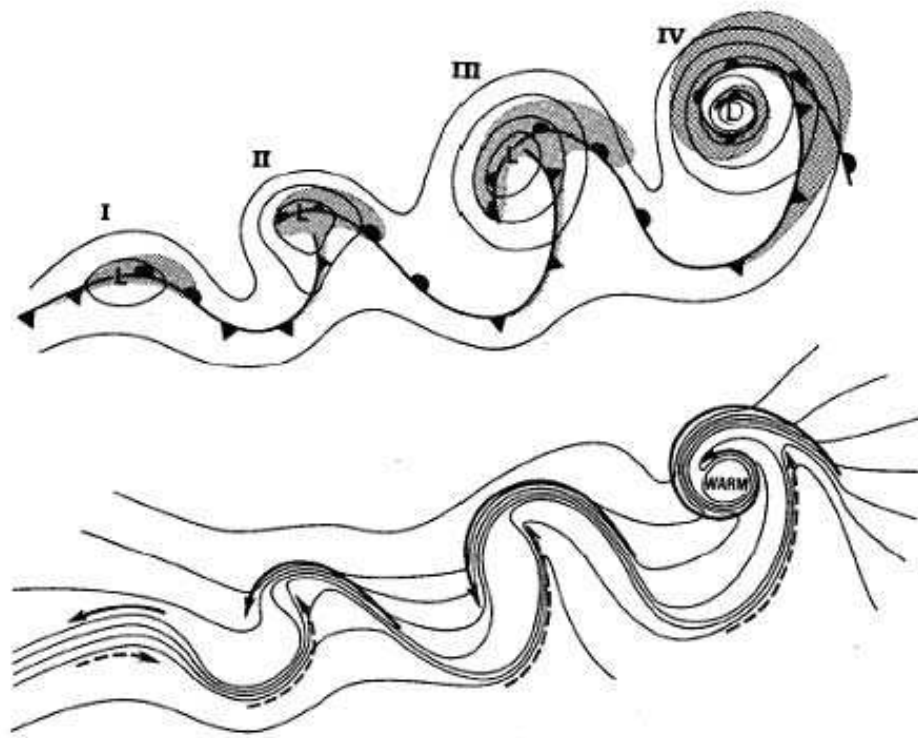
Figure 2.14: Shapiro-Keyser (1990) frontal-cyclone evolution: incipient broad-baroclinic phase (I), frontal fracture (II), bent-back front and frontal T-bone (III), and warm-core frontal seclusion (IV), Upper: sea level pressure (solid), fronts (bold), and cloud signature (shaded). Lower: temperature (solid), and cold and warm air currents (solid and dashed arrows, respectively).

A feature of the Shapiro-Keyser model is the enclosure of a cylinder of relatively warm pre-frontal air by advected cold air at the centre of the vortex – termed the *warm seclusion*. In contrast to the Norwegian model, no lifted occlusion is present. Lifted occlusions are not unknown, but are much less common than had previously been assumed. Lifted occlusions are most likely to occur over land, rather than open ocean, and may be a secondary feature induced by the land surface topography.

Thorncroft, Hoskins and McIntyre (1993) examined the air flows associated with Polar-front cyclonic centres (fig.2.15). Warm air ascends from south of the Polar front and cold air descends from north of the Polar front. Both flows are able to diverge into anti-clockwise (*cyclonic*) and clockwise (*anti-cyclonic*) rotating components. It is suggested that the Norwegian model for mid-latitude cyclone development occurs when the *anti-cyclonic* flows *A* and *D* are strong, whereas the Keyser-Shapiro model is followed when the *cyclonic* flows *B* and *C* dominate.
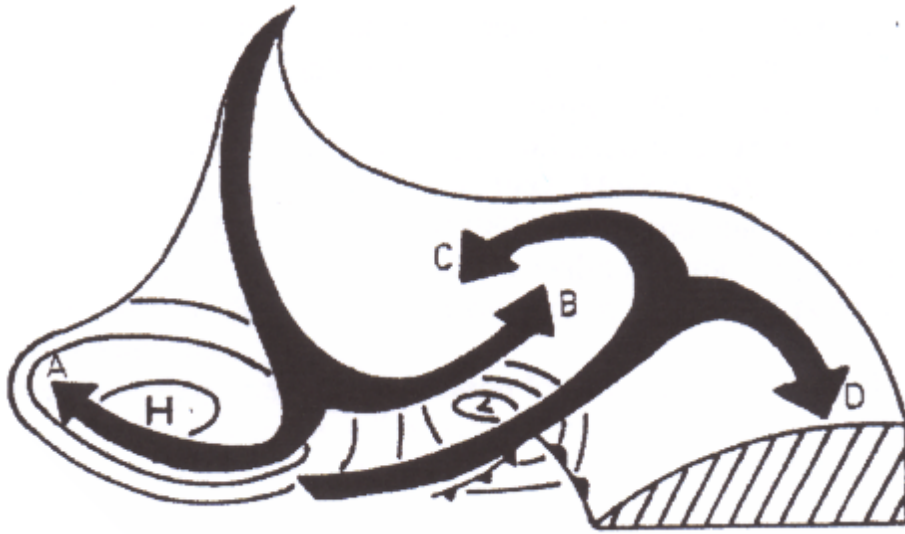
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Figure 2.15: Three-dimensional diagram looking northwards, illustrating the air flows associated with a cyclonic centre and trailing high pressure region. Warm air ascends from the warm sector and may diverge as flows C and D. Cold Polar air descends southwards and may diverge as flows A and B. (After: Thorncroft, Hoskins and McIntyre,1993)

Schultz, Keyser and Bosart (1998), and Schultz and Wernli (2001), have been able to make a further link between the various theoretical models by showing that the pattern of mid-latitude cyclone development can be related to upper air flow patterns. The *jet streaks* of the Polar Jetstream exhibit divergent air flows in left-hand exit regions and convergent air flows in right-hand exit regions. When a cyclonic centre comes under the influence of jet stream divergence, it tends to develop with a Norwegian pattern of fronts. However, if it comes under the influence of jet stream convergence then its development follows the Keyser-Shapiro pattern.

The example charts for 29 October 2005 (fig.2.16) illustrate a mid-latitude cyclone tracking north eastwards towards Britain. This is under the influence of the divergent flow in the left exit zone of the overlying jet streak, and is developing a Norwegian structure of occluded front:
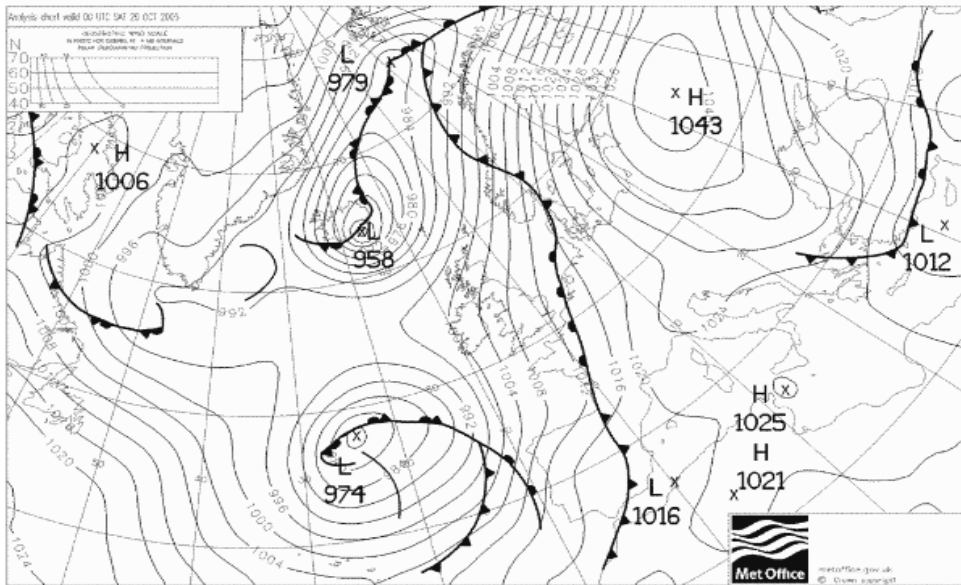
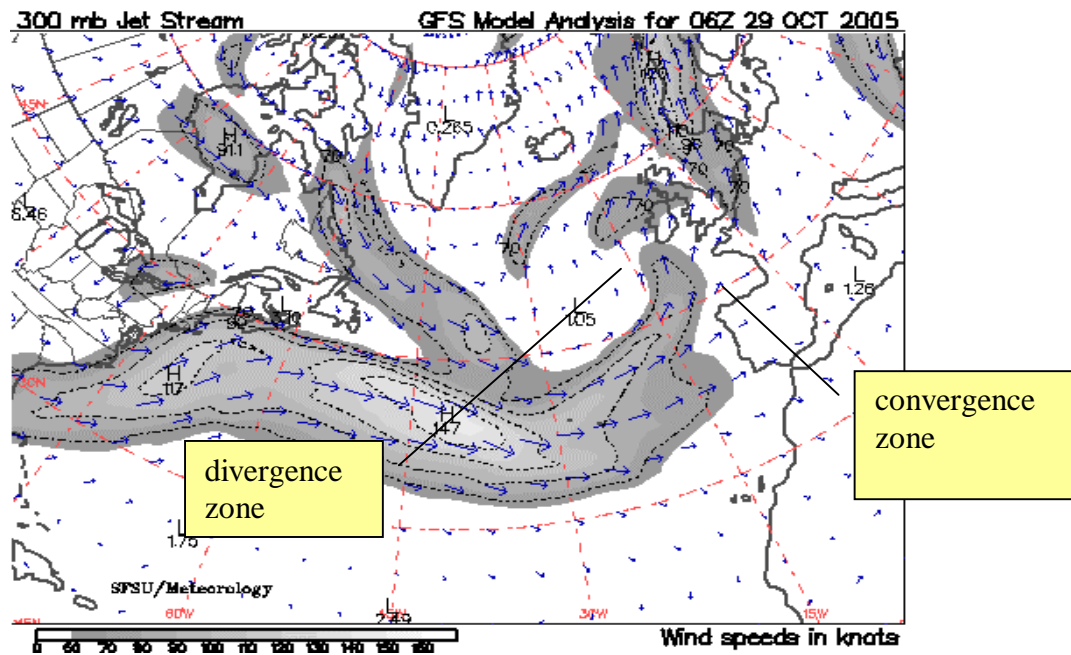**Figure 2.16(a): Synoptic chart for 00:00h 29 October 2005.** Source: Wetterzentrale.

Figure 2.16(b): Jetstream map for 06:00h 29 October 2005. Source: California Regional Weather Server

Rainfall patterns associated with Polar Front cyclones

Considerable interest has been focussed on the mechanisms of rainfall generation within extratropical cyclones, leading to the 'conveyor' model of Harold (1973). In this model, it is assumed that upwards displacement of warm air by the advancing cold front leads to convective instability. This is promoted by a mid-altitude flow of colder air crossing the warm sector, which maintains potential instability conditions as the warm air rises along an inclined surface termed the *conveyor* (fig.2.17). Initially the conveyor rises parallel to the cold front, but after crossing the warm front at mid-altitude it may swing round to join the flow direction of the high-level jet stream. A counter-current of cold air descends ahead of the warm front. This model, based on field data from rainfall radar and radiosonde balloon ascents, is consistent with the rainfall and air current patterns of the Shapiro-Keyser cyclone model (cf. fig.2.14).
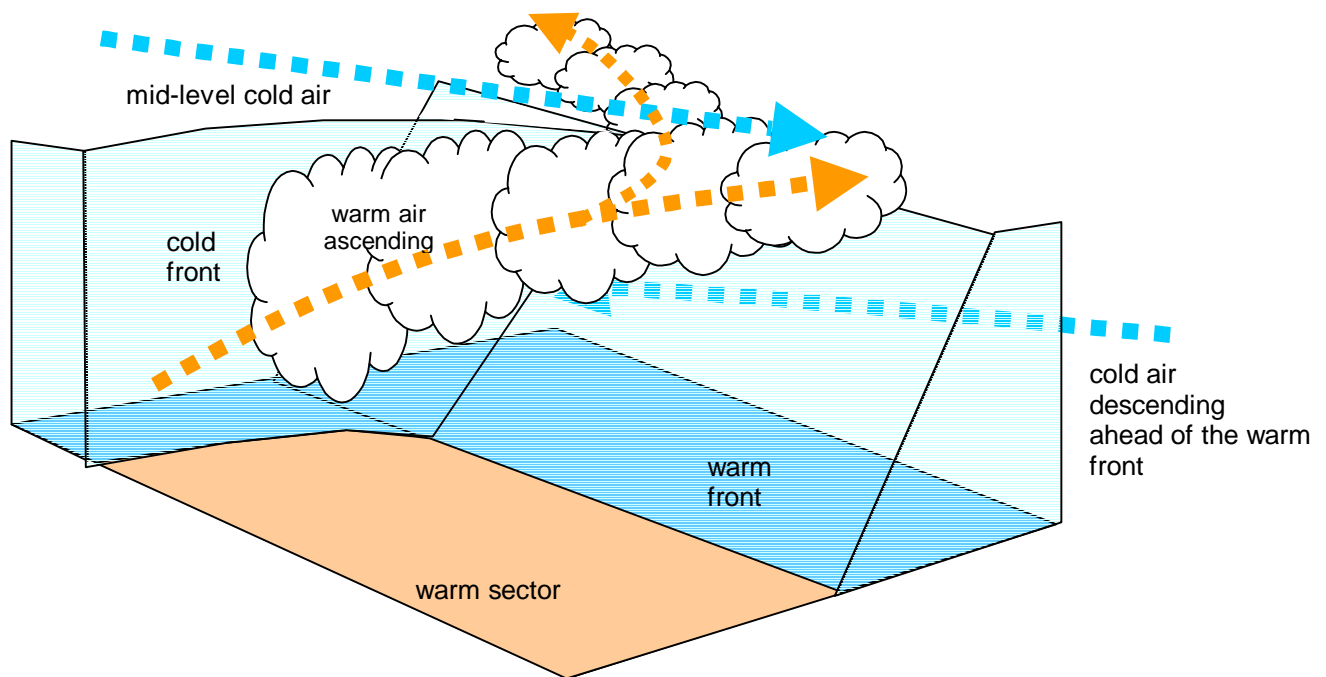
Figure 2.17: Ascent of warm air conveyor within a frontal system

A figure from the paper of Harold (1973) is reproduced here as fig.2.18. Values of θ_w shown in the diagram refer to *wet bulb potential temperature*. This is a measure of the temperature that an air parcel would take if brought to sea level without transfer of heat energy to or from the surrounding environment. Potential temperature is a useful concept, as it allows the comparison of temperatures of air masses without the

additional complication of computing temperature changes due to expansion or contraction during ascent or descent. It provides a measure of the relative quantities of thermal energy at different positions within a weather system.
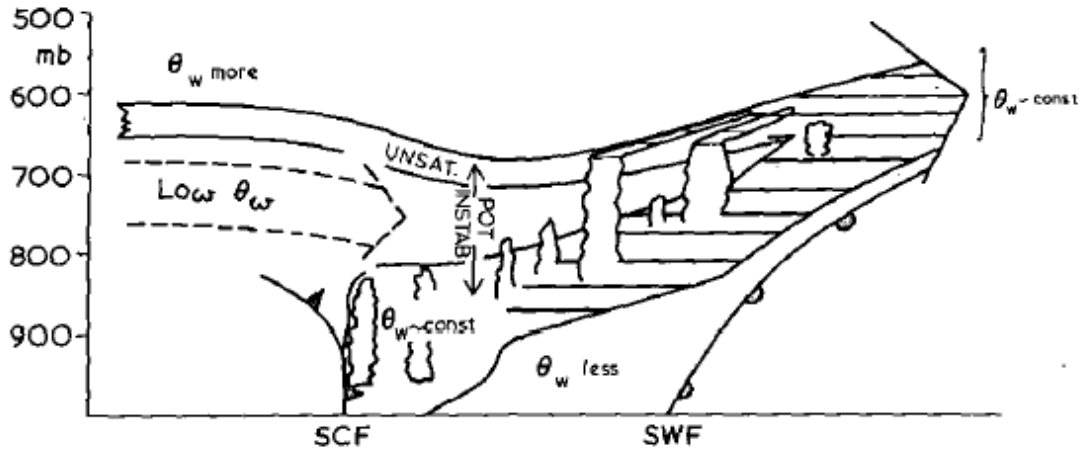

Figure 2.18: Schematic representation of the flows in the region of the cold front. Hatching denotes layer cloud and convective overturning is depicted by the cumuli form shapes. Two initially separate flows, with lower θ_w between, mix to produce a lapse of near constant θ_w above the warm frontal zone. Figure and caption reproduced from Harold (1973).

Browning, Hardman, Harold and Pardoe (1973) present a study of a frontal rainfall event over the Scilly Isles on 18 January 1971 which was intensively monitored by a combination of ground observations, rainfall radar, aircraft data collection and radiosonde balloon ascents. This generated an extremely detailed three dimensional model of the frontal system and associated rainfall as it approached Cornwall from the Atlantic. A series of rainbands were mapped, running parallel to the cold front and merging into a zone of uniform precipitation ahead of the advancing surface warm front (fig.2.19). These rainbands are seen as zones of convective uplift above the ascending warm sector conveyor. The intervening rain-free zones may represent the return flows of the convection cells, where descending air is warmed and re-evaporation of moisture takes place. The cause of the convective instability was again identified as a tongue of cold air over-running warm sector moist air ahead of the cold front (fig.2.20).

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Figure 2.19: Schematic distribution of precipitation [above the Scilly Isles for the rainfall event of 18 January 1971]. The extent of surface rain is shown stippled. The further extent of precipitation aloft is lightly stippled. The leading and rear edges of the dense high-level cirrus canopy is indicated by hatched shading. Figure and caption reproduced from Browning, Hardman, Harold and Pardoe (1973).

Figure 2.20: Schematic time-height cross-section representing some of the principal features of the convective rainbands on 18 January 1971. Cloud areas are stippled. Figure and caption reproduced from Browning, Hardman, Harold and Pardoe (1973).

Browning and Hill (1985) give a model for the relative positions and orientations of the Polar Front jet stream at high altitude, and the lower level ascending warm air conveyor associated with a cyclonic low pressure centre situated within the Polar air mass (fig.2.21):
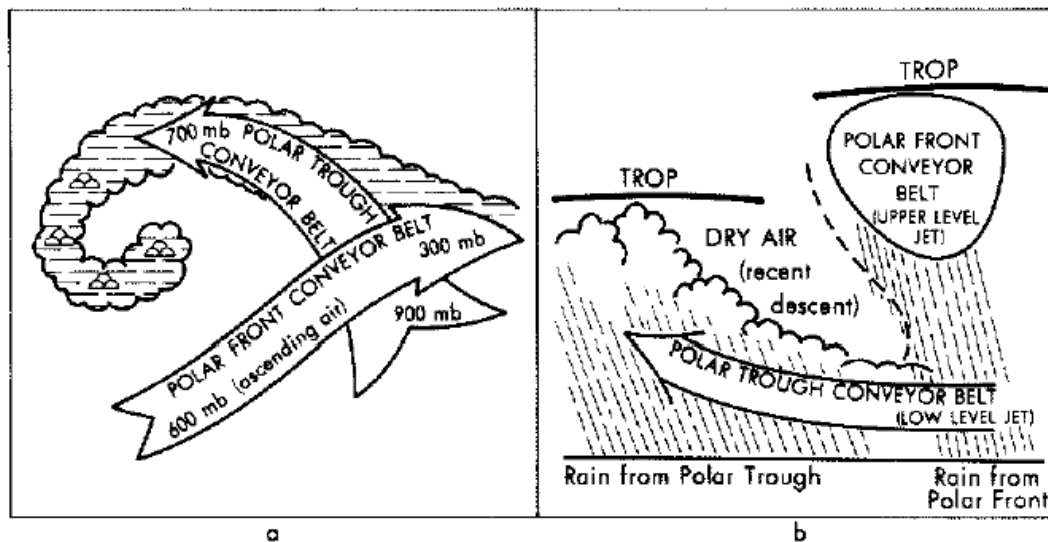

Figure 2.21: Conceptual model showing intersecting polar trough conveyor belt and polar front conveyor belt: (a) plan view, (b) vertical section along axis of polar trough. Figure and caption reproduced from Browning and Hill (1985).

An example of the approach of a cold front to the coast of Cardigan Bay is illustrated in fig.2.22. A middle-altitude conveyor flows northwards, with a well developed cumulo-stratus cloud layer indicating condensation from moist ascending air. Convective instability is producing rounded cloud tops. At higher altitude the conveyor will swing eastwards to join the track of the jet stream, picked out here by trails of cirrus ice cloud.

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Figure 2.22: Cold front approaching the Rhinog mountains.
C: mid-altitude conveyor   J: high-altitude jet stream. Photo: John Mason

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Orographic effects on rainfall

The production of rainfall discussed so far has been on a regional scale, controlled by the location of synoptic features such as the jet stream, low pressure centres and frontal systems. For detailed prediction of rainfall over an upland catchment such as the Mawddach, modification of rainfall patterns by smaller mesoscale processes must be considered.

The lowest 1000m of the troposphere is termed the planetary boundary layer. Localised effects are produced in this zone, including frictional drag from vegetation, deflection of winds around hills, convection from ground heated by solar radiation, or convection from warm sea water. These effects need to be considered in a high resolution meteorological model.

Fluid dynamic processes for airflow over hills are discussed by Raupach and Finnigan (1997). They identify three air flow regions:

- *Inner region* on the upflow side of a hill where the air flow pattern is influenced by both a pressure gradient due to the hill obstruction, and also a frictional effect due to the land surface.
- *Outer region* forming the higher air layer on the upflow side of the hill, where only pressure effects are experienced.
- *Wake region* on the downflow side of the hill where turbulence may occur as a result of flow separation.

Barry (1992) discusses the factors which affect whether an air flow ascends over a hill obstruction or is deflected around the obstruction (fig.2.23). The critical height h_c at which a streamline within the airflow is deflected *around* a hill rather than *over* it is given by the equation:

$$h_c = H(1 - F)$$

where H is the height of the hill.

F is the Froude number in the range [0,1] which represents a ratio of viscous forces to gravitational forces. In the case of air flow approaching a hill, this may be interpreted as the ratio of the kinetic energy of the air flow to the potential energy needed to rise over the hill.

There are two limiting cases:

- For fast moving air approaching a low hill, the Froude number will be close to 1, resulting in h_c near to zero. Almost the full depth of the airflow down to ground level will be able to surmount the hill.

- Slowly moving air approaching a high barrier will generate a Froude number close to zero. The critical height h_c will be approximately equal to the height of the hill, H. Almost the entire airflow below the hill summit will be deflected around the obstruction.
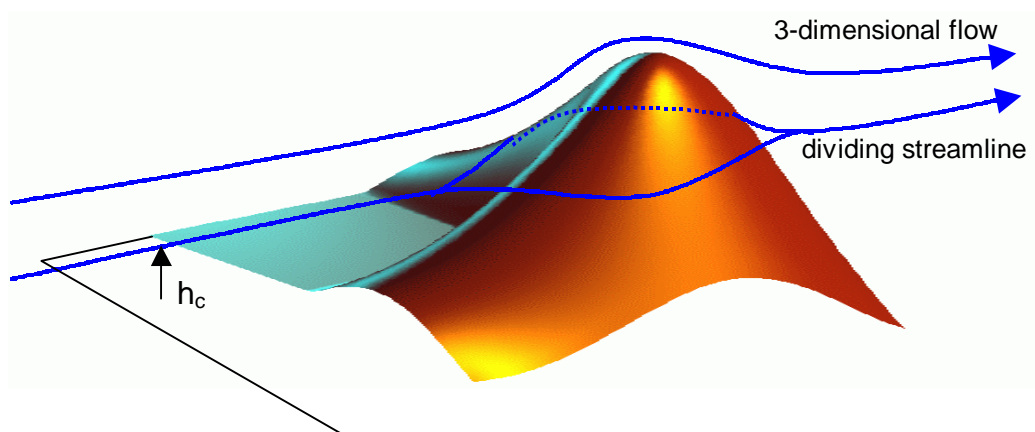

Figure 2.23: Flow paths around an isolated hill of height H

Airflow ascent over mountains can lead to cooling and condensation. If the air mass is in a stable buoyancy state, it may descend again beyond the mountain barrier and reabsorption of water vapour can occur. This situation is illustrated in fig.2.24, where a stable air flow is generating cloud as it streams north-westwards over the Aran mountains.

Chen and Lin (2003) have carried out modelling to investigate precipitation associated with unstable airflow over mountains. They identify different cases in which rainfall may occur upstream of the mountain summit, over the mountain summit itself, or over the downstream slopes. Critical factors are: the air flow velocity which affects rate of forced ascent and cooling, and the convective available potential energy (CAPE) of the air mass which affects the rate of free ascent after upwards displacement.

If atmospheric conditions are neutral in their buoyancy effect, a series of standing waves known as *lee waves* can be generated downstream of the mountain barrier. The waves are maintained by air circulation in rotors above the ground surface (fig.2.25).
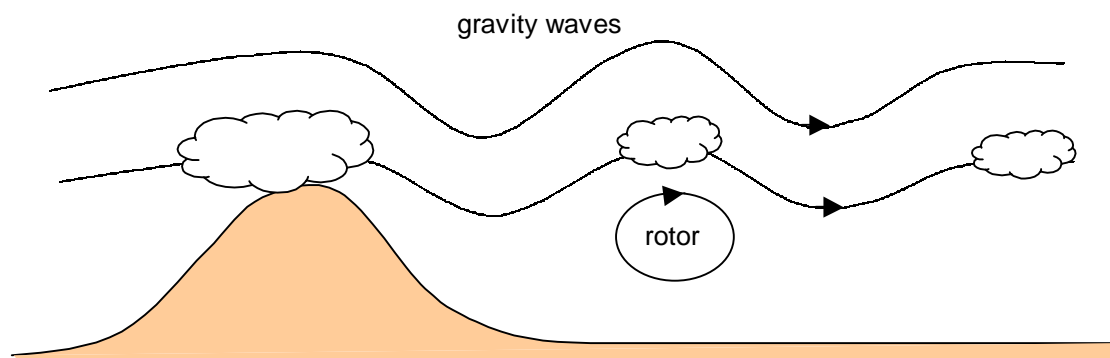

Figure 2.25: Development of lee waves (after Barry and Chorley, 1976)

Galvin and Owens (2006) have presented a satellite photograph showing the development of lee waves over the British mountains on 25 February 2006 (reproduced in fig.2.26). Lines of cloud developed in response to a stable airflow from the east. Lee wave clouds appear where the airflow crosses the northern Pennines towards the Lake District, and cloud ridges of longer wavelength can also be seen over Wales.
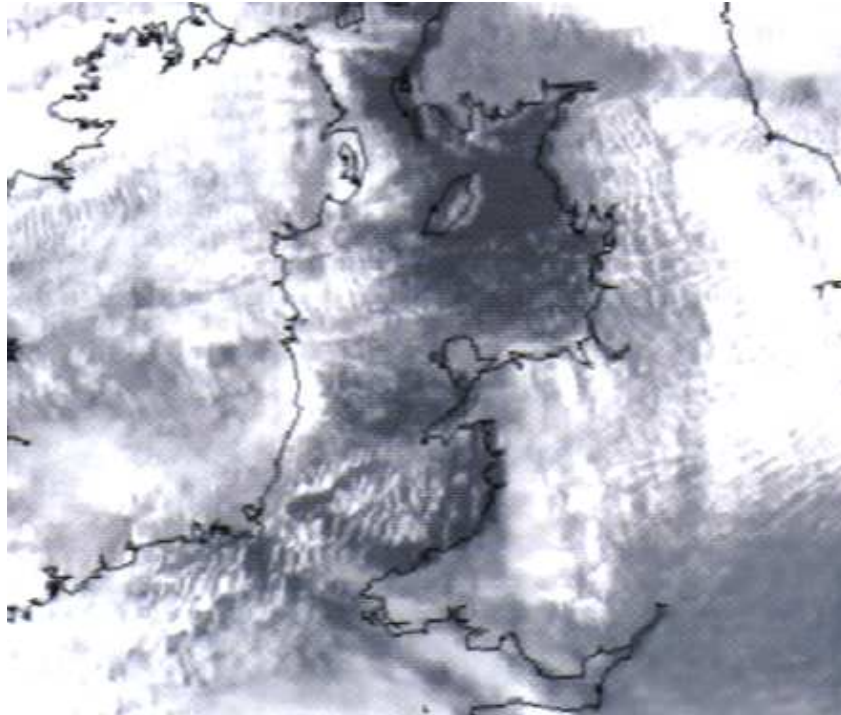
Figure 2.26: Development of lee waves over the Welsh mountains and the Lake District on 25 February 2006. From Galvin and Owens (2006)

Where atmospheric conditions are unstable, or neutral in their buoyancy effects, air given an upwards motion by passage over a mountain barrier may continue to rise. Condensation may lead to growth of water droplets within cloud, and the release of rainfall.

Garvert, Colle and Mass (2005) collected data from aircraft and radar to show that gravity waves over the Cascade Mountains of Oregon produced enhanced rainfall in the vicinity of narrow mountain ridges. The authors went on to simulate the storm events using the MM5 meteorological model and found that rainfall was increased by 20% in comparison to a smooth slope model without localised gravity wave forcing.

Rainfall generation over mountains may occur by a two stage process termed the *seeder-feeder* mechanism (Sibley, 2005). Cloud layering is commonly present over mountains during rainfall events, with low level orographic cloud formed by advection over the mountain barrier and higher level cloud produced by uplift along a warm air conveyor. This situation is illustrated in the photograph of the Rhinog mountains in fig.2.22. Small diameter rain droplets descend from the upper *seeder* cloud, then accrete further moisture as they pass through the lower *feeder* cloud. The resulting rainfall may be two or more

times the intensity of frontal rainfall alone. Strong low level winds ahead of the front may advect the falling raindrops some distance beyond the mountain barrier, so that the location of maximum rainfall does not correspond exactly with the area of highest surface altitude (fig.2.27).

upper level seeder cloud

lower level feeder cloud

advection of raindrops by airflow across the mountain barrier
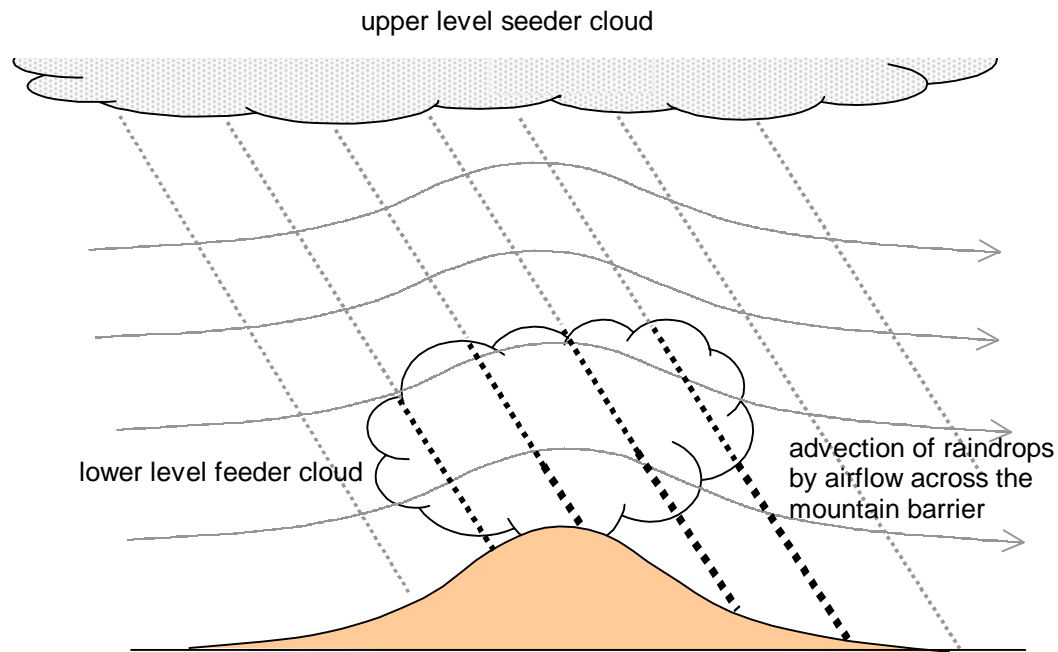
Bader and Roach (1977) develop formulae for estimating the amount of water vapour washed out by droplets falling through low level orographically produced cloud.

Convective storms

The most intense storm events recorded in the Mawddach catchment are associated with extremely unstable warm air masses undergoing convective uplift. Where a rising air parcel continues to be warmer than the surrounding air, uplift can extend to great altitude. Atmospheric temperatures near the tropopause are well below the freezing point of water so ice crystals separate by freezing, rather than the condensation of water droplets. Cool, descending counter-currents can develop in clouds of great vertical extent, leading to convective cycling and the accretion of ice crystals into larger hail particles. In the process, static electrical charges can be separated and conditions develop for the discharge of lightening. A typical thunder cloud structure is illustrated in fig.2.28. Particular features often present are an upper cloud extension, termed an *anvil*, close to the tropopause, and a low level cloud development ahead of the main convective structure, termed a *roll cloud* or *shelf cloud*. A *gust front* may occur below the leading edge of the thunder cloud, with high winds of varying direction causing damage at the ground surface. Precipitation typically commences as hail, with a progression to heavy rain as the storm advances.
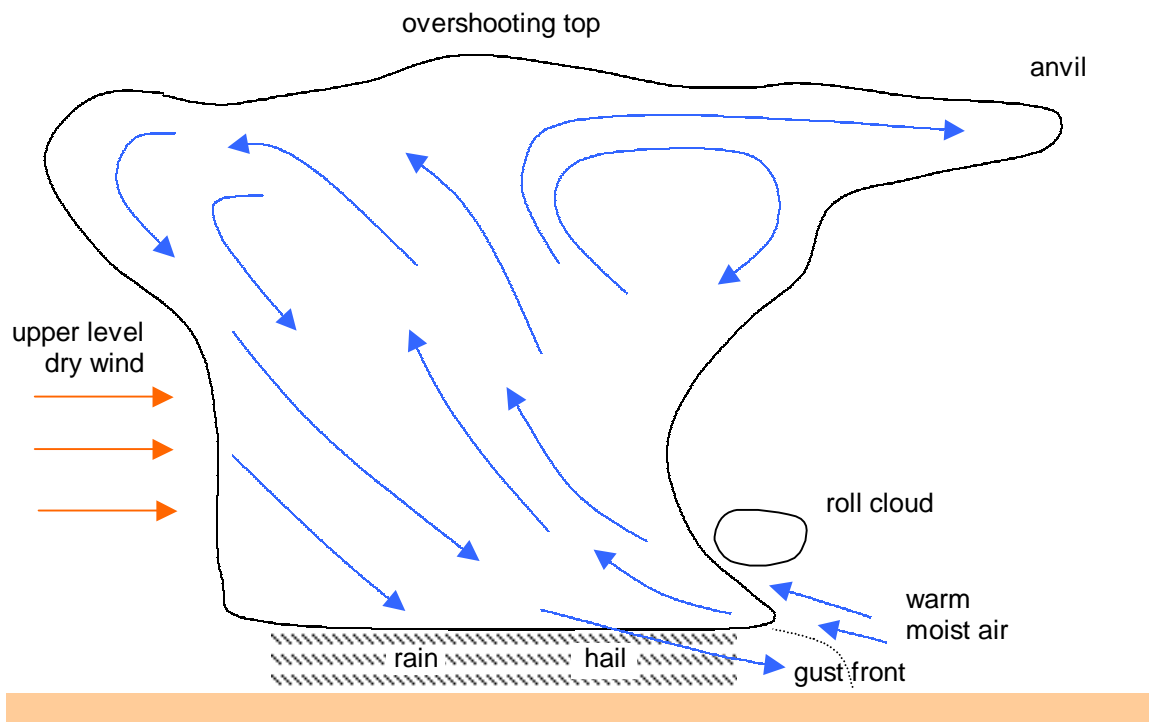

Figure 2.28: Structure of a typical thunder cloud (after Short, 2006)

Squall line structure

Thunder clouds may develop in isolation, but the most destructive storms occur when a line of thunder clouds are formed in association with a synoptic scale front. This structure is known as a *squall line* (National Weather Service, 2005). The storm of 3 July 2001 which caused widespread damage in the Mawddach catchment was of this type.

A section of squall line over the Midwest of USA is shown in the rainfall radar image fig.2.29 (National Weather Service, 2005). Red colour shading indicates zones of maximum rainfall intensity. Embedded within the squall line is a *bow echo* structure – a curved line of thunderstorms associated with strong vertical convection. Cold air is drawn in at the rear of the bow, forming an inflow notch of lower rainfall intensity. The most powerful thunderstorm activity is often just to the north of the bow apex, with the development of strong cyclonic vortices carrying warm air aloft.
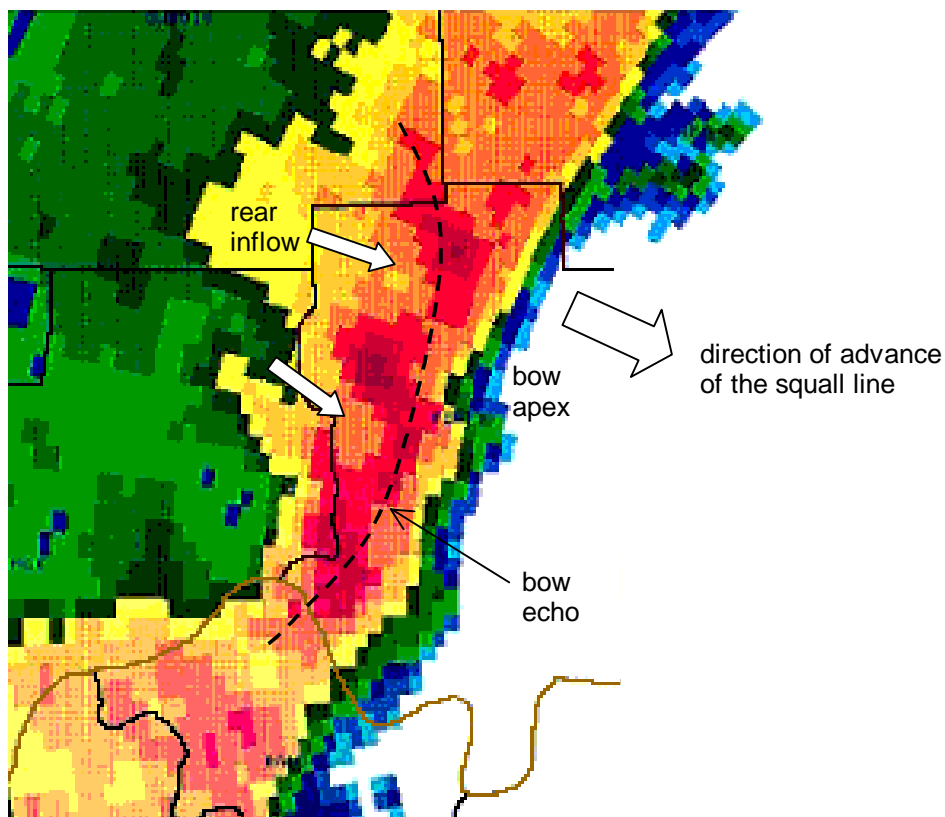

Figure 2.29: Rainfall radar image of a squall line over Kentucky
source: US National Weather Service

The leading edge of the squall line generally produces a sharp rainfall boundary. There may, however, be a large area of stratiform precipitation extending to the rear of the main rainfall band due to high level air flow from the front to the rear of the squall line.

Several theories have been proposed for the development of thunderstorm squall lines ahead of a cold front, and may variously account for different storm events. The cold front may push forward, so that cold air aloft intrudes above warmer air near the ground. This situation leads to severe instability. Intrusion of a cold air wedge at ground level ahead of the main front can impart an upwards motion to the warm air mass, which will lead to continued convective uplift if instability is present. Subsiding cold air behind the squall line can complete the return flow for the convective cell and drive further uplift (fig.2.30).
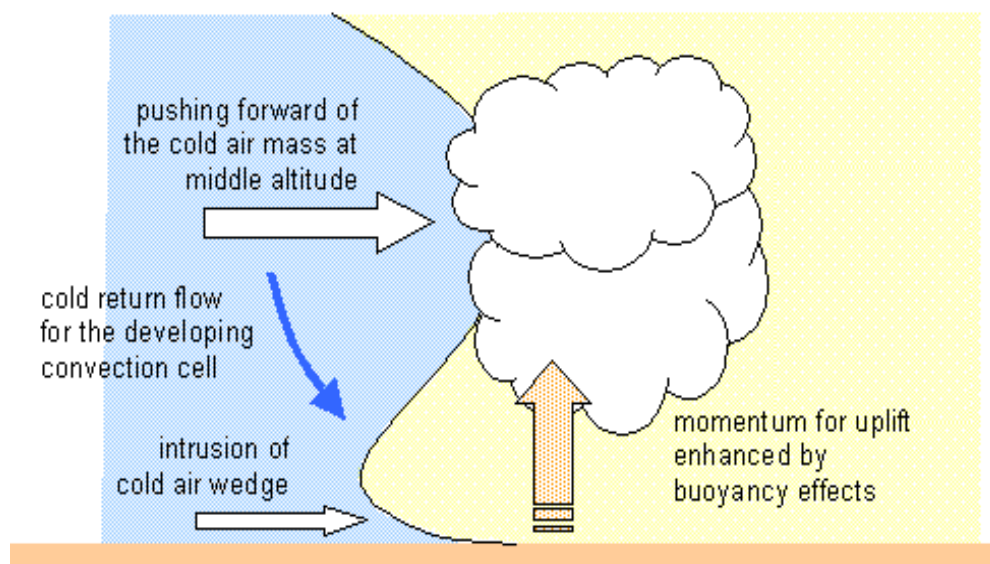

Figure 2.30: Convective uplift ahead of a cold front.

Mechanisms for squall line development have been examined in detail by Fovell and Tan (1998, 2000), illustrated here as fig.2.31:

A body of air sufficiently warmer than its surroundings can begin to ascend through buoyancy forces. The ascending air body compresses the air above itself, causing an increase in pressure. Pressure is relieved by the

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establishment of contercurrents flowing downwards around the margins of the ascending air body. In this way a convective cell is established (fig. 2.31a).
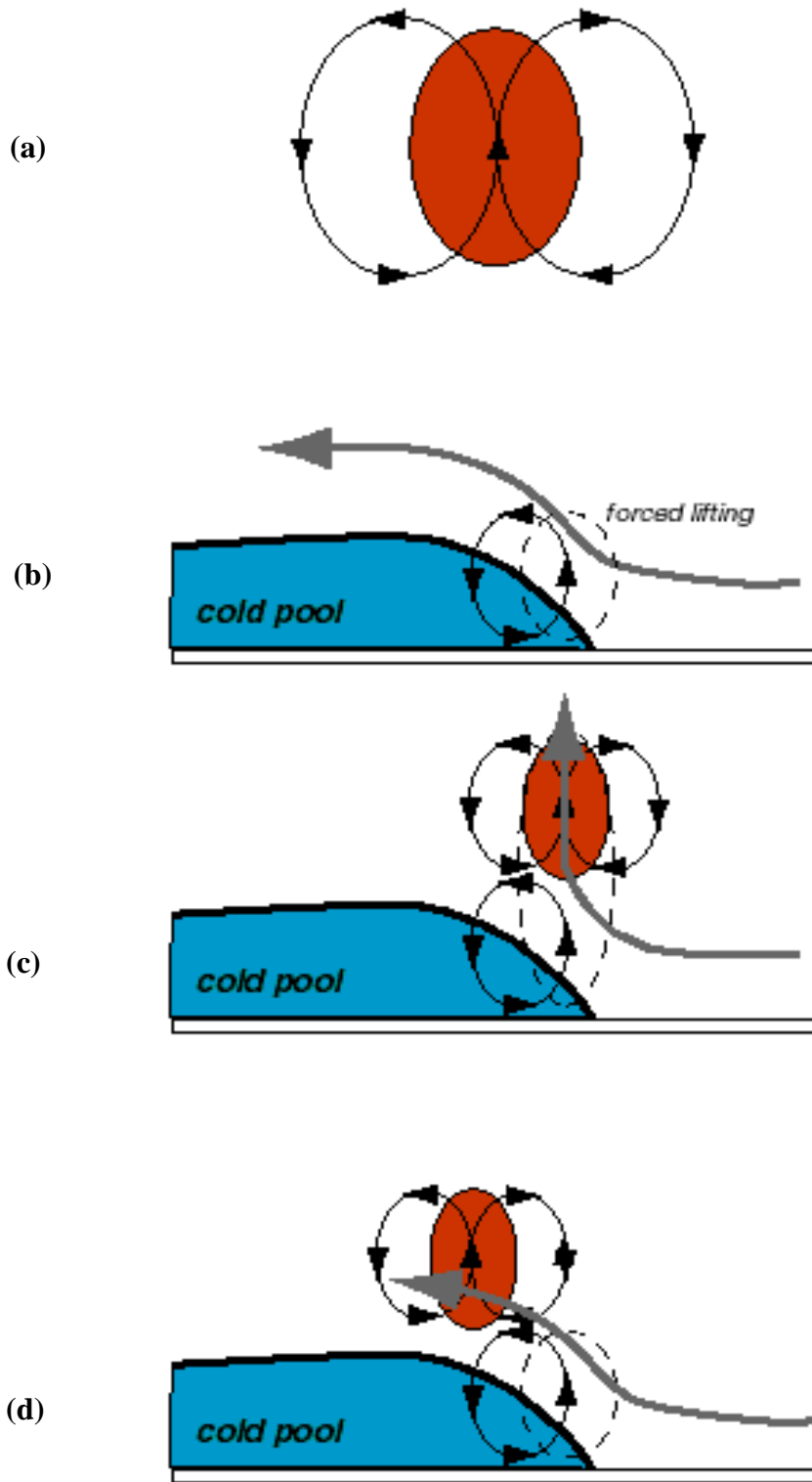

(a)

(b)

(c)

(d)

Figure 2.31: Diagrammatic representation of convective processes at a squall line. After: Fovell and Tan, 1998.

Fovell and Tan assume convective activity at a squall line is initiated by the extension of a tongue of cold air at ground level ahead of a cold front. This advancing cold pool displaces warm moist air upwards. An asymmetric convection is established, with a countercurrent of cooler air descending rearwards into the cold tongue (fig. 2.31b).

If suffcent instability is present, bodies of warm air may begin to rise at the front of the cold tongue (fig. 2.31c). Interaction with the low level circulation causes the convective cells to be advected backwards above the cold pool at the same time as they are developing in vertical extent (fig. 2.31d). Cold air begins to enter the lower parts of the convection cell, and a point is reached where convective instability is removed. This results in the final fragmentation of the convection cell and upwards motion ceases.

The theory developed by Fovell and Tan (1998, 2000) can explain many charcteristics of squall lines, including: their position some tens of kilometers ahead of an advancing cold front, their narrow lateral extent, and the well developed flow of cold decending air at the rear. The authors have tested their theory by field observations and detailed modelling. An example model sequence is shown in fig.2.32. This figure combines displays of potential temperature (shown by shading, with dashed contours for a negative temparature zone representing the low level cold pool), and vertical air velocities (shown by solid contours for upwards motion and dotted contours for downwards motion).

Frame A illustrated the initiation of convective activity in a body of warm air displaced upwards by the advancing.

In frames B and C, the convection cell is advected backwards above the cold tongue and reaches maturity. The cell develops to a great vertical height, and shows a strong downwards countercurrent at the rear. Rainfall would be intense at this stage due to convective cooling and condensation within the ascending air flow.

In frame D, cold air at mid-troposphere levels is mixing with the decending air flow and stabilising the convective cell. Temperatures begin to fall and the cell fragments, bringing convective activity to a close. Simultaneously, however, a new

convective cell is initiated at the forward edge of the cold tongue and the sequence begins again.
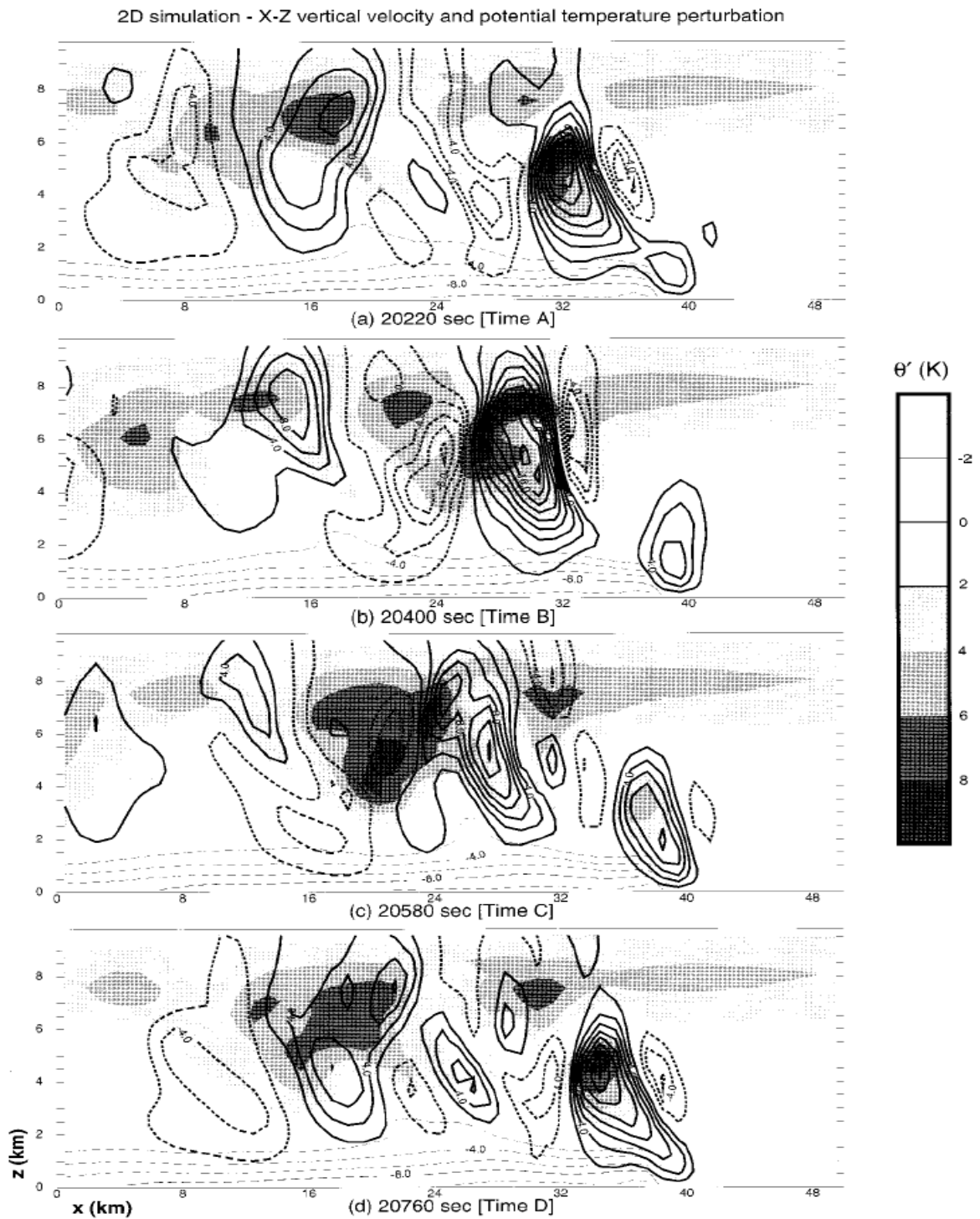

Figure 2.32: Example modelling sequence showing a cycle of convective cell development within a squall line. From: Fovell and Tan (1998).

Summary

- Most rainfall in North Wales is associated with mid-latitude cyclones which have crossed the Atlantic.

- Cyclones move eastwards roughly following the track of the Polar Front and its associated Jet Stream.

- Cyclones may also develop in isolation within the Polar air mass north of the Polar Front.

- Cyclones associated with the Polar Front have two main development patterns. Cyclones following the Norwegian model evolve to produce an occlusion of warm air undercut and lifted from the surface by cold Polar air. Cyclones following the Shapiro-Keyser model evolve to produce a cylindrical warm-core seclusion surrounded by colder air.

- Linking between a cyclone and the overlying Polar jetsteam can determine whether development will follow the Norwegian or Shapiro-Keyser model.

- Rainfall associated with a mid-latitude cyclone is produced by ascending air flow within the warm sector, termed the 'conveyor' and marked by cloud development.

- On making landfall, mid-latitude cyclones may encounter mountains which promote rainfall by forcing the uplift of moist air to the level of condensation.

- Mountains may induce gravity waves in neutrally stable air, or may initiate the continued uplift of unstable air.

- Raindrops originating from high cloud layers may grow by accretion of water vapour during their descent through low level cloud, known as the *seeder-feeder* mechanism.

- Some of the most severe floods in North Wales are caused by convective processes, where unstable warm moist air rises to produce thunderstorms.

- Thunderclouds may align along a cold front to produce a squall line. The squall line may remain almost stationary for a period, with convective cells repeatedly growing and decaying along its length.

- All of the processes listed above need to be incorporated in a successful rainfall prediction model for the Mawddach catchment. It will be shown below that the MM5 meteorological model can represent all these processes to a high degree of accuracy, with the possible exception of some convective thunderstorm events.