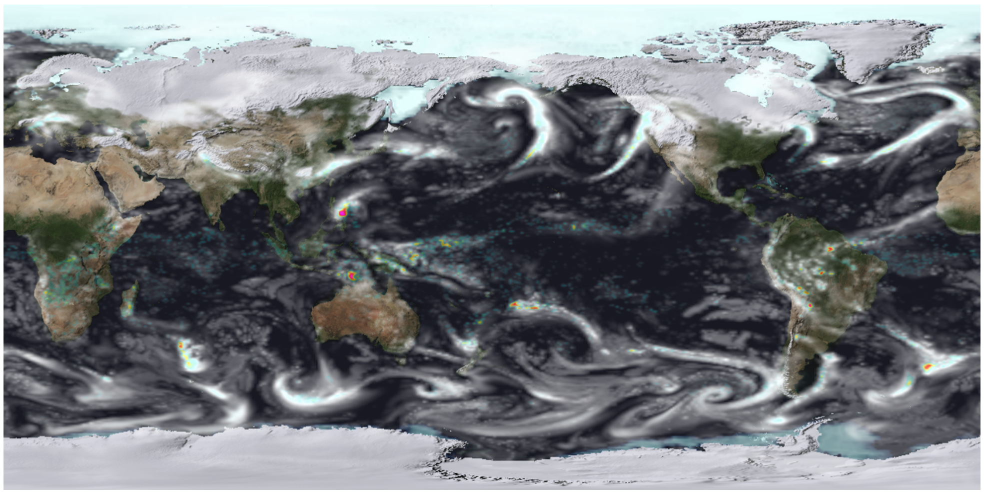
NCAS Climate Modelling School 2015



Climate Laboratory notes — Flat Earth

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Chapter 1

Introduction

1.1 Motivation

The relevance of mountains in determining the atmospheric flow and climate on Earth is convincingly motivated in the introduction of Smith (1979):

"It is often said that if the Earth were greatly reduced in size while maintaining its shape, it would be smoother than a billiard ball. From this viewpoint the mountains on our planet seem insignificant, and it makes us wonder how they manage to have such a strong influence on our wind and weather. One answer to the dilemma is that the atmosphere itself is very shallow — a density scale height of about 8.5 km — so that many mountains reach to a significant fraction of its depth. This argument, however, underestimates the mountain effect. The real answer is that our atmosphere is exceedingly sensitive to vertical motion — and for two reasons.

First, its strong stable stratification gives the atmosphere a resistance to vertical displacement. Buoyancy forces will try to return vertically displaced air parcels to their equilibrium level even if such restoration requires a broad horizontal excursion or the generation of strong winds. Second, the lower atmosphere is usually so rich in water vapor that slight adiabatic ascent will bring the air to saturation, leading to condensation and possibly precipitation. As an example, the disturbance caused by a 500-m high mountain (i.e., a very small fraction of the atmospheric depth) could well include (a) broad horizontal excursions of the wind as it tries to go around rather than over the mountain, (b) severe downslope winds as air that has climbed the mountain runs down the lee side, and (c) torrential orographic rain on the windward slopes. . . . "

For some well-defined idealised situations, it is possible to obtain analytical solutions of the governing equations describing the atmospheric flow over and around orography — a number of examples are provided in Smith (1979). In deriving these solutions, assumptions have to be made concerning

- the nature of the fluid (e.g., incompressible, Boussinesq, barotropic, or baroclinic fluid)
- the shape, and horizontal and vertical extension of the topographic barrier
- boundary conditions such as the upstream velocity profile, the stratitfication, and the char-

acter of the upper boundary

One of the main factors determining the flow regime is the mountain width and how it compares with several natural length scales in the atmospheric system including (with scale increasing):

1. the thickness of the amospheric boundary layer
2. the distance of downwind drift during a buoyancy oscillation
3. the distance of downwind drift during the formation and fallout of precipitation
4. the distance of downwind drift during one rotation of the Earth
5. the Earth radius

Analytical studies have provided invalueable insight into how the atmosphere is influenced by mountains. A comprehensive treatment of more realistic situations, including the vastly complicated forcing and dissipation in the troposphere, requires the use of numerical models. A difficulty that arises when analysing numerical simulations is that the models are so complicated that the model output, just like the atmosphere itself, cannot always be readily explained from first principles. Nonetheless, modelling studies offer the possibilty to conduct controlled quantitative experiments helping to narrow the gap between theory and observations. In this lab, we will conduct one such experiment where the role of mountains in Earth climate will be illustrated by comparing with a hypothetical "flat Earth".

1.2 Recommended literature

A plethora of studies investigating the role mountains in the climate system have been conducted. A small selection of these studies is provided in the reading material accompanying this lab.

Manabe and Terpstra (1974) explain the importance of mountains in the general circulation of the atmosphere. In the upper troposphere and stratosphere, the presence of mountains affects strongly the stationary flow fields, creating troughs over large mountain areas and modifying the surrounding circulation, partly due to an enhanced vertical transport of planetary wave energy from the troposphere to the stratosphere (Kasahara *et al.*, 1973). In the lower atmosphere, the presence of orography has a strong impact on the kinetic energy of stationary and transient disturbances. The probability of cyclogenesis is also modified, as are hydrological processes such as the nature of moisture transport, leading to changes in the global distribution of precipitation.

In a more regional study, Kitoh (2004) shows the effects of mountain uplift on the East Asian summer climate, experimenting with different mountain heights using a coupled atmosphere-ocean model. Systematic changes in precipitation and circulation appear with progressive mountain uplift, and sea surface temperatures are affected via feedback processes. Orlanski and Gross (1994) study lee cyclogenesis at a west-east oriented mountain range. A sensitivity study removing the East African Highlands has beend conducted by Slingo *et al.* (2005) and shows a substantial impact on the climate of Africa, India, Southeast Asia, and the Indian Ocean. The role of land-sea contrast and orography, in particular the Rocky Mountains, for the location and intensity of the

North Atlantic storm track is investigated by Brayshaw *et al.* (2009) using a hierarchy of idealized and 'semirealistic' simulations with the HadAM3 model.

Potential vorticity conservation is a powerful concept that can help to interpret atmospheric flow over topography. An introduction following Holton (2004) is part of this document; for more detailed information concerning the interpretation of potential vorticity maps see, e.g., Hoskins *et al.* (1985).

The review article by Smith (1979) has been introduced above.

1.3 Objectives of this exercise

In this exercise, you will be using a **general circulation model** (GCM) of the Earth's atmosphere to investigate the influence of continents, and especially large mountains, on the global circulation of the atmosphere. The GCM you will be using is the fast Earth-system Model FAMOUS Smith *et al.* (2008), which is one piece of the model hierarchy known as the **Unified Model (UM).** This model is an **coupled atmosphere-ocean** GCM with a dynamic carbon cycle. Your version of FAMOUS will be run at horizontal atmospheric resolution of N24 — a regular grid of 7.5° in longitude by 5° in latitude — and with 11 levels in the vertical. The ocean component uses a 3.75° longitude x 2.5° latitude grid with 20 levels.

You will use this model to investigate the role of mountains for the general circulation and climate on Earth. Some theoretical background is introduced in section 2. If you have a background in atmospheric physics and dynamics then most of this material will be familiar to you, but you may still find it useful to read before analysing the simulations.

However, this knowledge is not needed to setup the experiment and we can decompose our objectives as follows:

1. To modify the settings of an existing Unified Model integration in the **Unified Model User Interface *(UMUI)*** to setup an experimental integration with flattened orography, as described in a separate document provided. This integration will be compared with a control integration with default values of the orography. In making these modifications, you will gain knowledge of the inputs required for a UM integration and the diagnostic outputs that the model can provide for your analysis.

2. While the model simulation completes, to formulate hypotheses of what changes in circulation and climate you expect to find in the flat-earth experiment. Use the material covered in this document, the summer school lectures, and the recommended literature. In fact, you may find it efficient to form a 'mini journal club' with your fellow 'flat-earthlers'.

3. To make use of post-processing and graphics tools to analyse the model's output data. You will be provided with examples and templates of common functions that you can mimic when you perform your own analysis.

4. To critically evaluate the limitations of this sensitivity experiment.

5. To prepare a short presentation summarising your work for the other participants of the School.

Chapter 2

Theoretical background

2.1 Hadley circulation

This part describes the global circulation of the atmosphere, at a very large scale, which is not especially due to continents (note that these processes would then be the same as on an aqua-planet - a water cover Earth - an idealised setup often used in models to understand basic aspects of model formulation and the global circulation).

Globally, Earth emits as much energy as it receives. But at regions smaller than the globe and over time periods of less than a year, there is a significant imbalance in the distribution of **radiative energy** at various latitudes: the tropics receive a surplus of energy, while the poles run a deficit. This imbalance creates a **temperature gradient** from the equator to the poles, and associated density and pressure differences, which create movement of air modified by the Earth's rotation.

In order to restore the latitudinal energy balance, the energy is moved away from the Tropics. This circulation, called the **Hadley Circulation** (Fig. 2.1), explains the strong interactions between energy and water transport, as well as the atmospheric circulation and the global distribution of temperature and precipitation.

At the equator, the warm and moist air masses coming from both hemispheres meet, creating a zone of convergence with low pressure. The buoyant air rises, cools and water vapour condenses to form clouds and eventually precipitation. At the tropics, this convergence zone is known as the **Intertropical Convergence Zone** (or ITCZ) and is known as being an area of heavy rain throughout the year. At the top of the troposphere, air moves towards the poles, becomes cooler, denser and sinks. As the dry air sinks, it warms, which prevents condensation from occurring and clouds from forming, creating a **zone of subsidence** with high pressure, clear skies and low rainfall amounts. This zone corresponds to the largest deserts in Australia, Arabia and Africa. Back at the surface, the air moves from zones of high pressure to low pressure, which closes the Hadley circulation.

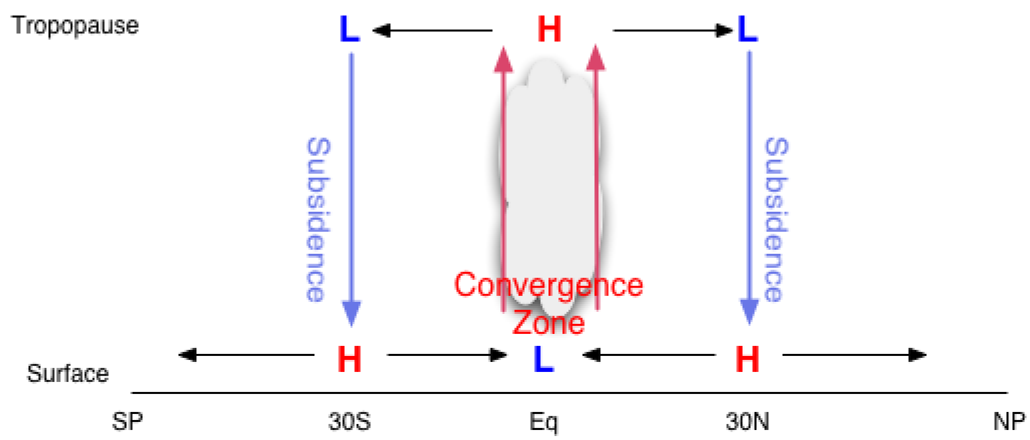

Figure 2.1: Simplified representation of the Hadley cell

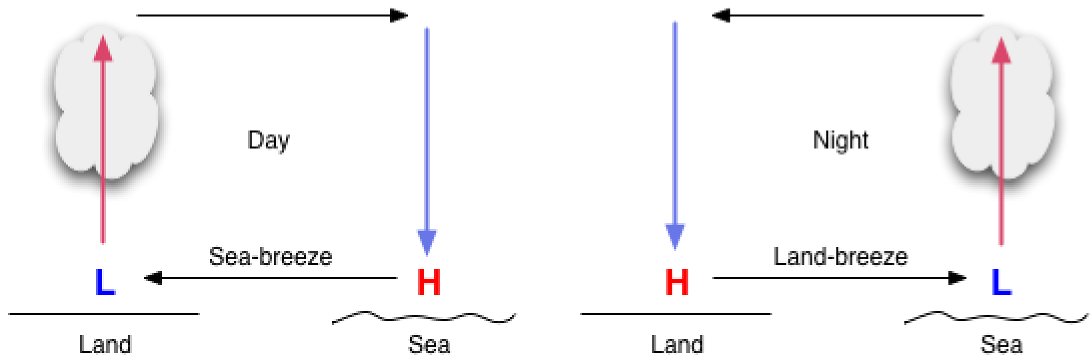

Figure 2.2: Schematic representation of the sea breeze.

2.2 Effects of continents

The Hadley cell is not continuous around the globe. It is affected by land-ocean contrasts in albedo, thermal properties and seasonal variability:

- The ocean has a smaller **albedo** than the land and therefore absorbs more solar radiation. Moreover, the land **thermal capacity** is lower than the ocean's so that surface temperature over land varies a lot more than over the ocean. Consequently, the diurnal cycle is stronger over land than over ocean. This temperature contrast modifies the circulation of air and creates sea or land breezes along coastlines as schematised in Fig. 2.2.

  During the day, air over land warms quicker than over ocean, creating a thermal contrast between land and ocean. This effect produces a pressure difference, with an air flowing from the ocean's high pressure to the land's low pressure zones. That is the sea-breeze observed along coastlines. Eventually at night, if the temperature over land cools down enough, the air becomes warmer over the ocean and the system reverses.

- The differences in land and ocean's thermal characteristics also have an effect on the circulation at larger timescales than a day. In fact, due to land's low thermal capacity, the tem-

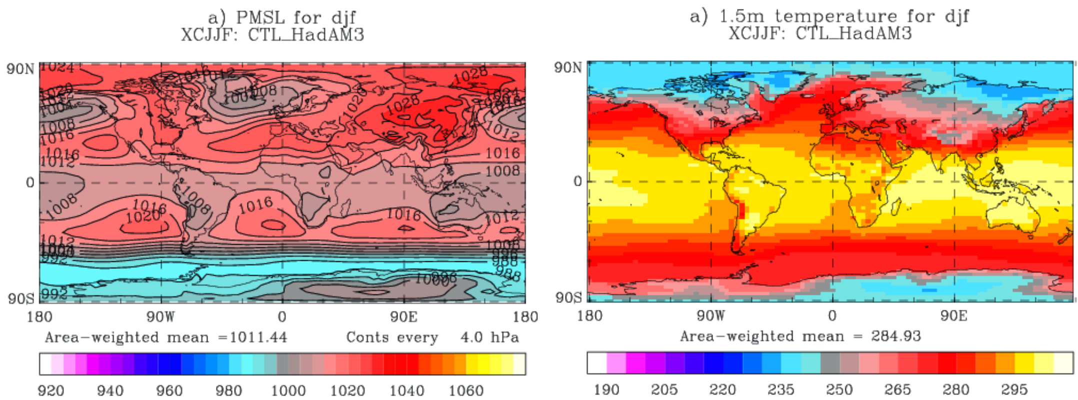

Figure 2.3: Mean sea level pressure (left) and temperature at 1.5m (right) for the control run in Dec-Jan-Feb.

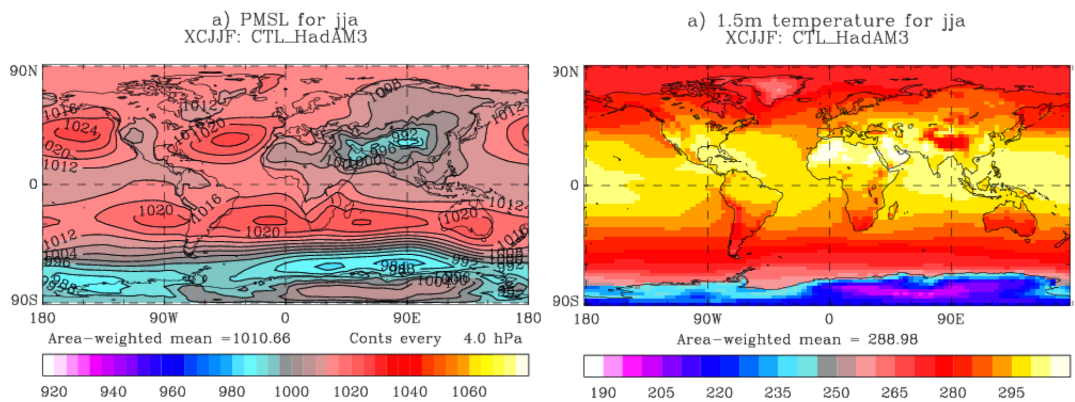

Figure 2.4: Mean sea level pressure (left) and temperature at 1.5m (right) for the control run in Jun-Jul-Aug.

perature difference between summer and winter is much larger over land than over ocean. This seasonal variability is especially large in the interior of large continental areas, far from the ocean. As shown in the figure 2.3, in winter the North American and Asian continents are cold, producing high-pressure zones (e.g. **Siberian High**). The North Atlantic and North Pacific have low pressure zones (**Aleutian Low**, **Icelandic Low**) associated with a steep temperature gradient between the mid-latitudes and polar regions. In summer (Fig. 2.4), the temperature gradient in the north mid- and high latitudes decreases, as well as the oceanic low-pressure zones, permitting the sub-tropical high-pressure zone to expand. Moreover, the continents become warmer with a low surface pressure. Note that the high and low pressure systems meet at zones of steep temperature gradients, creating an air masses difference and weather fronts. The southern hemisphere, dominated by ocean, shows weaker zonal temperature contrasts and a more zonal flow, which illustrates the large role of continents in the global circulation.

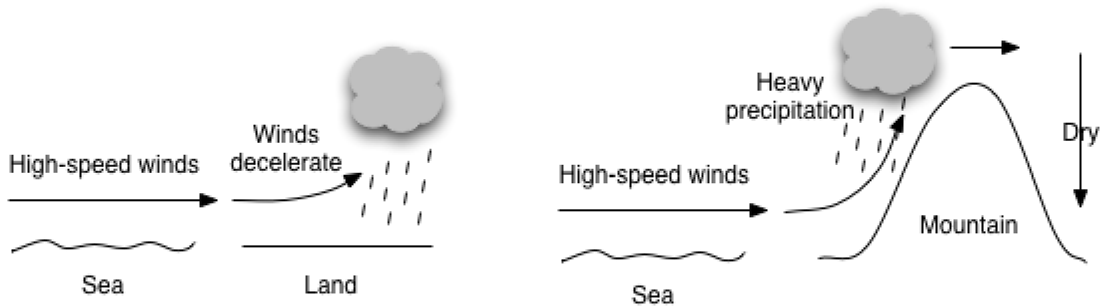

Figure 2.5: Effect of surface roughness on vertical air motion.

2.3 Effects of mountains

2.3.1 On vertical air motion

When the air coming from the sea meets land, it decelerates because of the **surface roughness** (this value varies depending on the surface cover: tree, grass, bare soil, city...). In the presence of a heating source at the ground (in summer for example), the moist air becomes buoyant and rises, creating a convergence zone with possible precipitation. If the flow encounters a jagged mountain, which has a much higher roughness, it is forced to decelerate strongly and move over the rising terrain (**orographic lift**). As it rises, it expands and cools adiabatically, creating clouds and heavy precipitation. A subsidence zone with clear sky often develops in the lee of the mountain, as schematised in Fig. 2.5.

Vertically propagating waves, caused by buoyancy force when the air is forced to rise, are called **internal gravity waves (or buoyancy waves)**. The formation and propagation of buoyancy waves depend on the stability of the surrounding air within the atmospheric stratified vertical layers (if the air was having to rise in unstable air, it would continue to rise without creating any wave pattern). It is possible for vertically propagating waves to be reflected, for example when they encounter a layer of strong vertical wind shear. In somes situations, the waves may be repeatedly reflected from an upper layer of air and from the surface downstream of the mountain. **Mountain lee waves** "trapped" in the lower troposphere are formed.

The introduction of a parametrization of **gravity-wave drag (GWD)** into GCMs had beneficial impact on the simulated circulation and temperature in the troposphere and stratosphere (Palmer *et al.*, 1986). We will see later that GWD in models depends on the subgrid-scale variability in topographic height, which is important to remember to build a Flat Earth.

Some flows may not be able to go over the mountain. An example is the circulation in the Indian Ocean in summer (Fig. 2.6; Slingo *et al.* (2005)). the easterly flows are blocked by the East African Highlands and turn, becoming south-westerly. This is an important mechanism for the Indian summer monsoon.

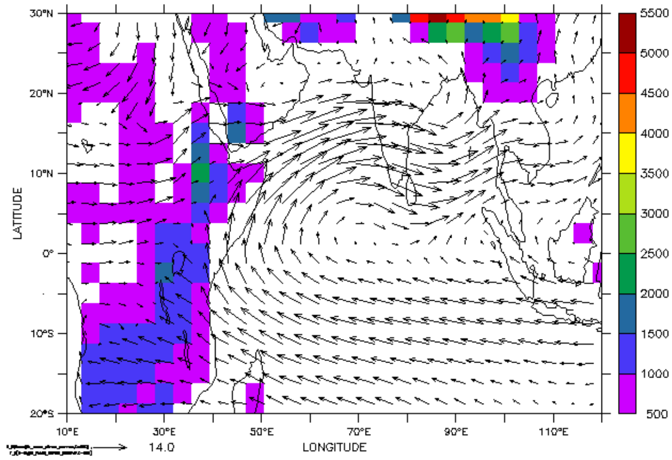

Figure 2.6: Orography and wind vectors at 850 hPa for the control run in JJA over the India Ocean.

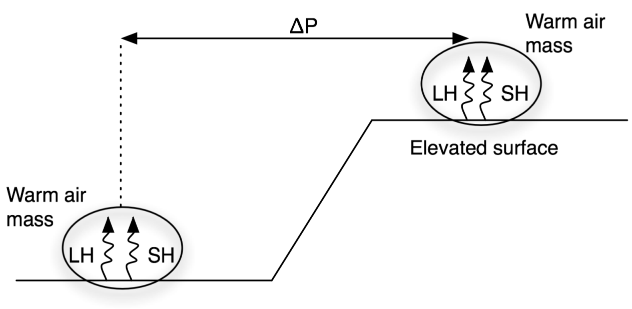

Figure 2.7: Mountain plateau acting as an elevated heat source.

2.3.2 On the distribution of temperature and pressure

The surface of a high-mountain plateau is a source of sensible heating (warming of the air by the land surface) and latent heating (condensational heating released in precipitation). This **elevated heat source** creates an upper-level horizontal thermal and pressure gradient as schematised Fig. 2.7. An example is the **Bolivian high**, an upper-level ($\sim$200 hPa) **orographic anticyclone** which develops during the summer over the high plateau region of the Central Andes.

2.3.3 On the circulation due to conservation of potential vorticity

The **potential vorticity conservation law** has been first introduced by Rossby, in order to define a field that would keep track of a flow and describe its evolution. It is especially useful to understand the generation of vorticity in cyclogenesis (birth and development of cyclones; Hoskins *et al.* (1985)) and to explain the effects of mountains on the large-scale flow.

The precise definition of PV depends on whether we consider a homogeneous incompressible fluid (constant density, $\nabla \cdot U = 0$), a barotropic fluid (in a region of uniform temperature distribution, $\rho = \rho(p)$, PV is called Rossby PV) or a baroclinic fluid (in a region of temperature gradients,

$\rho = \rho(p, T)$, PV is called Ertel PV). However, among all definitions, PV is always a measure of the ratio of the absolute vorticity η to the depth of the vortex h. For atmospheric large-scale flows, $\eta = f + \xi$ is the vertical component of absolute vorticity and controls the way the flow evolves, through the two following terms:

- The **planetary vorticity** $f = 2\Omega \sin\phi$ (strictly its vertical component, also called the Coriolis parameter), due to the Earth's rotation. By definition, f increases (decreases) when the flow goes poleward (equatorward).
- The **relative vorticity** $\xi = \frac{\partial v}{\partial x} - \frac{\partial u}{\partial y}$ (again, strictly its vertical component) due to the rotation of an air mass, relative to a frame rotating with the Earth. By definition, positive (negative) values of ξ are associated with cyclones (anticyclones) in the northern hemisphere and anticyclones (cyclones) in the southern hemisphere.

If we consider the simpler case of a homogeneous incompressible fluid (in shallow water system for example) in adiabatic motions, conservation of PV implies:

$$\frac{D}{Dt}\left(\frac{\xi + f}{h}\right) = 0 \quad \text{or} \quad \frac{\xi + f}{h} = \text{constant}$$

where $\frac{DX}{Dt} = \frac{\partial X}{\partial t} + U \cdot \nabla X$ refers to the change of a vector X following trajectories.

This law implies that if the fluid flows over a topographic barrier, as its column squashes (when approaching the obstacle) and stretches (after the obstacle), the vorticity fields ξ and f have to change and deviate the flow from its original track. If the depth of the fluid parcel is constant, then conservation of PV reduces to the conservation of absolute vorticity. However, if the depth of the parcel changes following the motion (as in the example below), it is potential vorticity that is conserved. In the two cases, westerly and easterly flows behave differently.

Example: we consider a westerly flow in the northern hemisphere. Initially, the flow is uniform ($\xi = 0$), as shown on the (x, z) figure below (top), which shows the behaviour of the vorticity fields. As it approaches the barrier, the flow near the surface will follow the contours of the barrier. Note that air higher in the atmosphere will also be deflected vertically, but due to pressure forces produced by interaction of the flow with the barrier, the deflection will be flattened and spread horizontally (see Smith, 1979, section 3.1 for details). Thus, the air column approaching the barrier will be stretched due to the upper-level deflection, compressed on the upslope, stretched again on the downslope, and then compressed towards their original depth downstream (Fig. 2.8).

Column stretching ahead of the barrier (h increases) causes a compensatory increase in $\xi + f$ so ξ becomes positive, leading to a northward cyclonic deviation. The change required in ξ is reduced by the increase in f as the parcel moves northwards, which makes the flow turn slowly (see Fig 2.8 in the (x, y) plane). As the parcel climbs the obstacle, column compression (h decreases) implies a decrease in $f + \xi$ for conserving PV. This generates negative anti-cyclonic relative vorticity, so the flow turns southwards. And as h continues to decay, the southward movement becomes stronger. On the downslope, column stretching again begins to turn the flow northwards. However, at the point downstream of the obstacle where the air column attains its original depth, the parcel is further south than it was originally, so that f is smaller and ξ is positive, implying a cyclonic rotation towards the pole, which increases f again but decreases ξ. When the parcel reaches its original latitude, it still has a northward component and continues poleward gradually, while ξ decreases until becoming negative. The parcel acquires an anti-cyclonic rotation and its direction reverses,

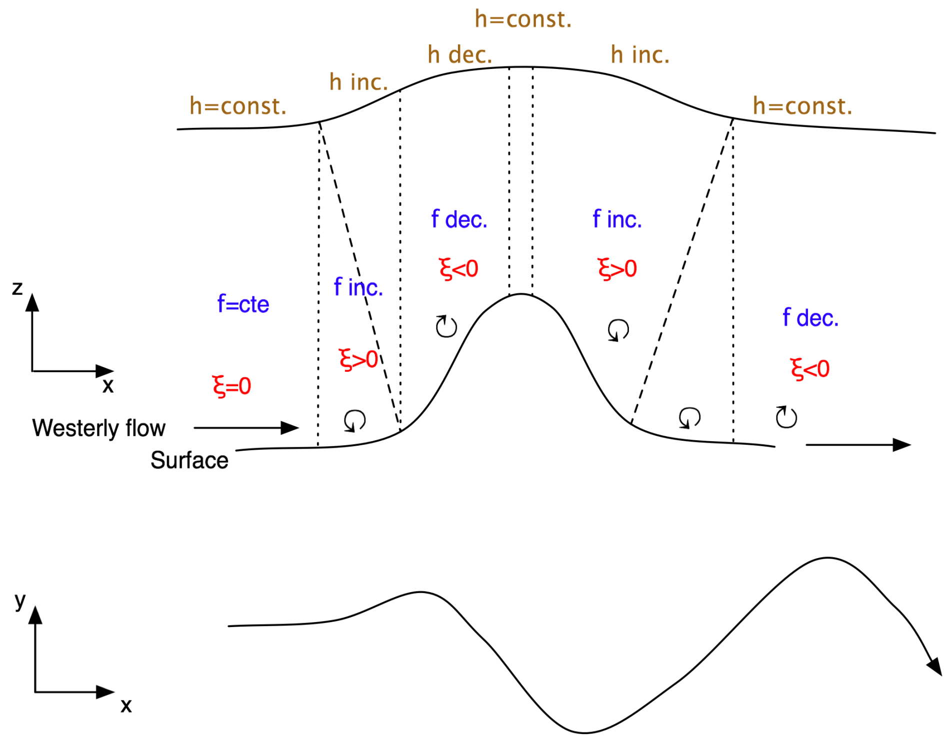

Figure 2.8: Schematic view of westerly flow over a topographic barrier (after Buzzi and Tibaldi, 1977). (top) depth of the fluid column as function of x, (bottom) trajectory of an air parcel in the (x, y)-plane.

moving southward. f decreases and ξ increases progressively until becoming positive, reversing its rotation and direction. Further downstream, the parcel will continue to create an alternating pattern of cyclonic and anticyclonic rotation, following a wave-like trajectory.

TASK: Using a similar analysis, it is possible to predict the effect of the barrier on an initially easterly flow. As an exercise, can you reproduce the behaviour of the vorticity fields (in a (x, z) plane) and the flow direction (in a (x, y) plane) in the blank space left below?

In this section, we only considered a flow perpendicular to the mountain. We did not discuss the theory of a flow aligned with the mountain. Orlanski and Gross (1994) make a good description of what happens in the Alps, and the consequences on cyclogenesis.

Chapter 3

Some ideas for analysis

- How does the midlatitude circulation change in the flat earth experiment? Have a look at surface pressure, and geopotential height and winds at different levels. See also Brayshaw *et al.* (2009).
- How is the Asian summer monsoon represented in the control and experiment runs? What is the role of topography over East Africa and the Tibetan Plateau? See also Slingo *et al.* (2005); Kitoh (2004).
- Do the meridional atmospheric transports of heat and moisture change? What about the ocean? See also Manabe and Terpstra (1974).
- Is the total mass of air the same in the control and experiment runs?

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