Arran field course 2019
Handbook

University of Reading

THE UNIVERSITY OF EDINBURGH

UNIVERSITY OF LEEDS

Contents

Overview

This is the ‘Arran Handbook.’ It contains the details of all the exercises to be undertaken on the field trip. It also contains detailed notes, explanations, information and general background to many aspects of atmospheric science.

The dominant focus of the course is the ways in which the atmosphere changes physically as you move through it in the vertical. We explore these changes on the scale of metres with a tower based at the field centre, on the scale of hundreds of metres by walking to the top of Goat Fell nearby and on the scale of kilometres by releasing radiosondes from the centre.

Another focus is the production of weather forecasts each evening for the next day and their subsequent analysis.

This handbook is a ‘work in progress.’ If you notice mistakes or have suggestions let a member of staff know. There is a copy of the handbook kept in the staff room to note mistakes and errors.

We hope you enjoy the module and gain from some of the more practical aspects of atmospheric science.

Jim McQuaid, Tristan Quaife & Richard Essery

Assessment

The module assessment is split across a number of activities, some of which include an “in field/class” element where your understanding/performance of the practical aspects of an activity and how you undertake it as a group is assessed. Reading MMet students have an additional component from a task completed at Reading, please see the module description for information.

Where applicable, yur infield assessment will be based on: your understanding of the exercise, observational skills, performance and ability to work as a team. After each interaction with staff you will be assessed on these skills and the average mark from all staff interactions will count towards your final mark.

An important aspect of the fieldcourse is working together as a team and time management, there is a lot of work to be done and you must ensure that you allocate sufficient time for each exercise. Your groups may subdivide for some of these exercises depending on the skills of the members of the group and the nature of the task.

For your notebook each of the assessed section (Weather forecasting, Radiosondes, Hill profiling, Surface Energy/Albedo/Turbulence and Balloon Profiling) carry equal weight within the module marks. Here you will be marked on your ability to follow the exercises and answer the questions posed in the handbook, but also for your ability to take the exercise further and to show a wider ability with the skills you have developed over your undergraduate career.

Your notebooks must be submitted at **9pm** on the final day of the fieldcourse.

Daily schedule

We expect a lot from you each day. The day will start with a weather forecast from a member of staff at 7:30am. At 8am there will be breakfast and activities commence at 9am. At the end of the evening each group will give a weather forecast starting at 9pm. Once they have all been completed you are free to relax (go to the pub if you wish). At this point the classroom will be locked and you will not be allowed to continue working on the activities.

Provisional Timetable for Arran2019

The timetable for the Arran trip needs to be relatively flexible to cope with changing weather conditions and other variables. Below is the provisional timetable for the fieldcourse this year.

Fri, 13th	09:00	11:00	LNCH	14:00	16:00	19:00	Sat, 14th	09:00	11:00	LNCH	14:00	16:00	19:00
Group 1	Lect	Lect	Lect	Walk	SP	class	Group 1	GF	GF	GF	GF	GF	class
Group 2	Lect	Lect	Lect	Walk	SP	class	Group 2	GF	GF	GF	GF	GF	class
Group 3	Lect	Lect	Lect	Walk	SP	class	Group 3	GF	GF	GF	GF	GF	class
Group 4	Lect	Lect	Lect	Walk	SP	class	Group 4	GF	GF	GF	GF	GF	class
Group 5	Lect	Lect	Lect	Walk	SP	class	Group 5	GF	GF	GF	GF	GF	class
Group 6	Lect	Lect	Lect	Walk	SP	class	Group 6	GF	GF	GF	GF	GF	class
MRes1	Lect	Lect	Lect	Walk	SP	class	MRes1	GF	GF	GF	GF	GF	class
MRes2	Lect	Lect	Lect	Walk	SP	class	MRes2	GF	GF	GF	GF	GF	class

Sun, 15th	09:00	11:00	LNCH	14:00	16:00	19:00	Mon, 16th	09:00	11:00	LNCH	14:00	16:00	19:00
Group 1	Hill	class		Pilot	Pilot	class	Group 1	ALB	SEE/SLD	Launch	SEE/SLD	Tephi	class
Group 2	ALB	class	Launch	Pilot	Pilot	class	Group 2	Tephi	Traj		class	Hill	class
Group 3	class	ALB		SEE/SLD	SEE/SLD	class	Group 3	Pilot	Pilot	Launch	Hill	Tephi	class
Group 4	class	Hill	Launch	ALB	Tephi	class	Group 4	Pilot	Pilot		SEE/SLD	SEE/SLD	class
Group 5	Pilot	Pilot	Launch	Hill	ALB	class	Group 5	class	Traj		Tephi	class	class
Group 6	Pilot	Pilot		class	Hill	class	Group 6	SEE/SLD	SEE/SLD	Launch	Tephi	ALB	class
MRes1	SEE/SLD	SEE/SLD		tephi	class	class	MRes1	Hill	ALB		class	class	class
MRes2	SEE/SLD	SEE/SLD		tephi	class	class	MRes2	Hill	ALB		class	class	class

Tue, 17th	09:00	11:00	LNCH	14:00	16:00	19:00	Wed, 18th	09:00	11:00	LNCH	14:00	16:00	19:00
Group 1	class	Traj		class	class	class	Group 1	class	class		class	class	class
Group 2	SEE/SLD	SEE/SLD		class	class	class	Group 2	class	class		class	class	class
Group 3	Traj	class		class	class	class	Group 3	class	class		class	class	class
Group 4	class	Traj		class	class	class	Group 4	class	class		class	class	class
Group 5	SEE/SLD	SEE/SLD		class	class	class	Group 5	class	class		class	class	class
Group 6	Traj	class		class	class	class	Group 6	class	class		class	class	class
MRes1	Pilot	Pilot	Launch	class	Traj	class	MRes1	class	class		class	class	class
MRes2	Pilot	Pilot	Launch	class	Traj	class	MRes2	class	class		class	class	class

Lunch is 13:00-14:00, groups rotate to assist with the 13:00 (local) radiosonde launch.

Additional sonde based on winner of competition to be launched either at 07:00 or 19:00 to be launched by winning team

Weather forecasting

Weather Forecasting

1. Aims of the exercise

Weather forecasting is a complex and difficult task that requires the synthesis of many different pieces of information, together with the experience and local knowledge of the forecaster, to produce a verifiable forecast. In this task we will try to simulate a real forecasting situation in a commercial forecasting company or on a scientific field measurement campaign that requires real-time forecasts of the next days' local and synoptic conditions. *The forecast period for this exercise is for the 24 hours between 6 pm on the day of the forecast and 6 pm the following day.*

HINT: always look outside just before 6 to see if it has been raining!!

Each day two different members of the group will have to produce two separate forecasts. A numerical forecast, which will be marked for its accuracy and a textual forecast which will be marked for the scientific reasoning which went into the forecast. Additionally, each day a different group will have to produce an oral forecast presentation which will be marked on the communication of the overall forecast situation.

The two members of each group making the forecast each evening should work together to produce each part of the forecast. The forecasting exercise should not take up too much of your time, and should be begun after dinner each evening. ***The deadline for textual and numerical forecasts to be submitted is 08:30pm***. There will be a short validation of the forecasts by a member of staff during the weather briefing the following morning.

You will have access to a number of different data sources to complete your forecast, which will be available from http://field1/ (bookmarked on your web browser). The following sections describe the requirements for your forecast, the data sources available and how to use them. Your group leader will give you guidance on how to interpret the data but you will need to decide which parts of the data are most important and what this means for the forecast.

2. Requirements of each forecast and marking scheme

Numerical forecast **35%**

You will be required to predict the following numerical values. The accuracy of your forecast will be assessed using data from the meteorological monitoring mast on the field site. You should take into account these local conditions when making your forecast (so go outside and have a look around!). Forecasts should be entered on the web form on the forecasting website. Regular updates of your accuracy will be given throughout the field course.

Forecast variable	Required Accuracy	Maximum score	Minimum score
Minimum temperature	0.5°C	4pts	-3pts
Maximum temperature	0.5°C	4pts	-3pts
Total rainfall	0.5 mm	5pts	-3pts
Wind direction @ 12z	10 degrees	4pts	-1pt
Wind speed @ 12z	1 ms^{-1}	4pts	-2pts

Textual forecast **35%**

A commentary on the development of the synoptic and local situation for the forecast period in no more than 200 words. You should include the dominant synoptic features, wind speed and direction and their changes, amounts of cloud and/or sunshine and timings of significant changes (e.g. onset or ending of rainfall). After the more detailed forecast for the next day, you should also provide a short outlook for the following 2–3 days. Use ensemble predictions to assess the uncertainty of the forecast. The most important part of the textual forecast is that you provide evidence for your conclusions from the data available and describe how you reached your conclusions (in this way, you textual forecast will hence look a little more technical than a newspaper/web forecast, because you will include justification/background).

Forecast presentation ***30%***

A presentation of no more than 6 minutes with no more than 6 slides and including no more than 3 images in total. The presentation should be in PowerPoint and be left on the computer server (you will be given details of how to do this) before the 7PM deadline. You will be awarded marks based on your ability to communicate the main elements of the forecast and your choice of graphics to illustrate your forecast. Try to interest the audience by linking the current weather to our outdoor activities. You will lose marks for exceeding the limits on time, number of slides and for including unnecessary information.

3. Features of synoptic-scale structures in mid-latitudes

There are several features of synoptic-scale flow in the mid-latitudes, which will be key to understanding the large-scale conditions, which influence the day-to-day weather in Arran. These will be most obvious on the Met Office analyses and forecast charts, but will have signatures in other diagnostics too. Figure 1 shows a range of these features, which we will examine in this section.

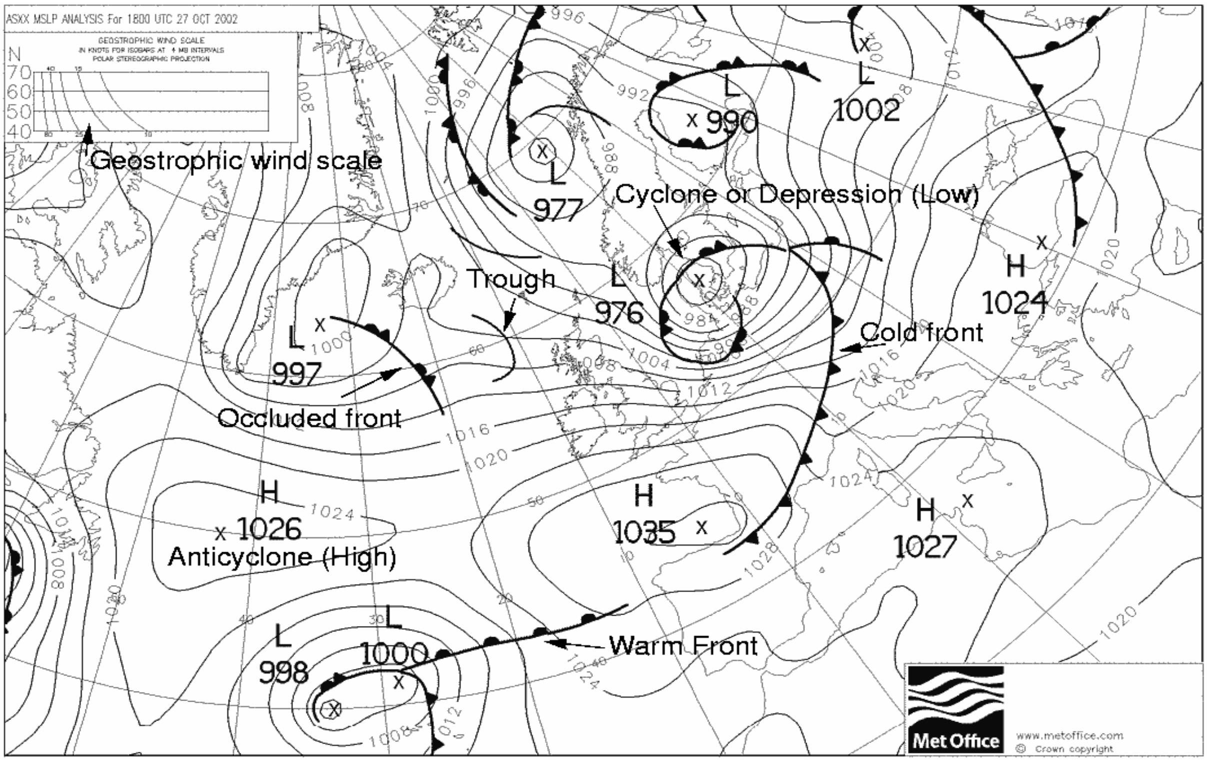


Figure 1: Typical synoptic features in the mid-latitudes. Taken from Met Office surface analysis for 1800GMT on 27 October 2002

3.1 Geostrophic flow

On horizontal scales of greater than 1000 km, mid-latitude atmospheric flow is approximately *geostrophic*. This means that air flows under a dominant force balance between the Coriolis force, due to the Earth's rotation, and the pressure gradient force. In a normal fluid flow, motion is from high pressure to low pressure but geostrophic balance changes this so that the flow is virtually along lines of constant pressure (isobars). Furthermore, the closer the isobars, the faster the flow. Hence from the isobars on a surface analysis chart we can immediately gain an impression of the wind speed and direction.

3.2 Thickness

Air pressure decreases with height, and so it is possible to measure the height difference between two levels above a point on the surface. Thickness, in synoptic meteorology, refers to the height difference between two standard pressure levels, 1000 (near the surface) and 500 hPa (usually around 5-6 km). Thickness is usually measured in units of tens of metres or decametres. The significance of the thickness measurement is that the colder the atmosphere the more dense it becomes and therefore the shallower the layer between 1000 and 500 hPa is. Thickness is a useful measure of the large-scale temperature through the lower part of the atmosphere. A useful rule of thumb is that if the thickness is less than 528 dm, then any precipitation is likely to fall as snow rather than rain. The angle between the thickness and surface pressure contours can also give an indication of the sign and amount of thermal advection.

3.3 Cyclones and Anticyclones

Regions of high surface pressure are called anticyclones or sometimes simply "highs." Regions of low surface pressure are called cyclones or depressions or sometimes "lows." In the northern hemisphere, the air flows clockwise around anticyclones and anticlockwise around cyclones. Cyclones are regions of large-scale ascent of air. As the air rises, it expands and its temperature falls. The lower the temperature, the less capacity the air has to hold water vapour and so condensation occurs, forming clouds. Hence cyclones are regions of cloud formation and rainfall. Conversely, anticyclones are regions of large-scale descent of air. Descending air is compressed and increases in temperature, increasing its capacity for retaining water vapour. Hence anticyclones are often (but not always) regions of clear skies and little rain. In winter there are certain conditions (particularly in which the anticyclone builds from the south-west) in which a great deal of shallow cloud or 'anticyclonic gloom' is associated with an anticyclone.

You will be able to see the cyclones and anticyclones on the surface pressure analysis and forecasts, and also in the satellite pictures which show the characteristic associated cloud patterns.

3.4 Fronts and Frontal Progression

These are surfaces, often only a few kilometres thick, between air masses of different properties. The term is particularly used to describe interfaces between air masses of different temperatures. Frontal surfaces are generally inclined at a small angle to the horizontal, with colder air under cutting warmer air. Their surface positions will be plotted on surface analysis charts but you will also be able to see the transition between warm and cold air masses in the radiosonde ascents which are made from Arran.

The direction of motion of the front determines whether, as it passes a fixed point, it marks a change to warmer or cooler conditions. There are three different types of frontal structure, warm fronts represent a transition from a cold to a warm air mass at the surface and cold fronts represent a transition from a warm to a cold air mass at the surface. The third type, occluded fronts, represent a more complicated structure where two or more frontal features have combined, they often have little surface structure but are still important because they are correlated with rainfall.

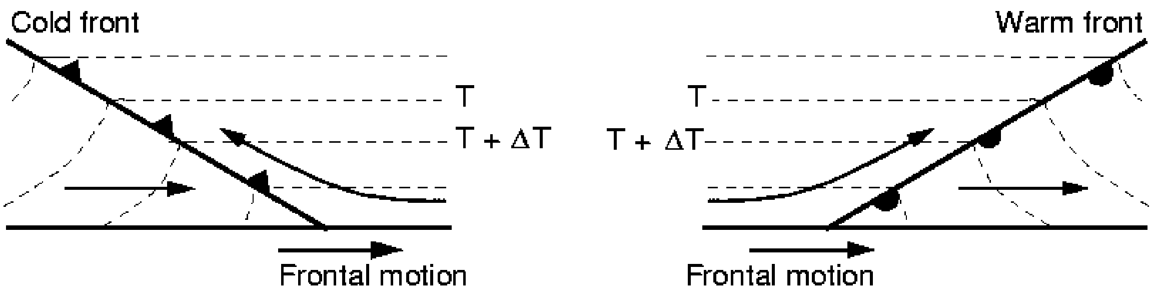


Figure 2: Schematic cross-sections through warm and cold fronts

Fronts are generated when horizontal convergence of the air flow brings surfaces of different air properties close together. Fig. 2 indicates air motion associated with fronts. For both warm and cold fronts, warm air is rising above cold air causing the warm air to cool as it expands, and hence cloud and rain formation. Thus fronts are potential regions of rainfall. Clouds tend to form along the frontal surface, with the cloud base close to the location of the front. Hence a typical frontal progression at a fixed site can often be observed of a gradual lowering or rising of the cloud base as the front passes by. Typical slopes are 1:50 for the cold front and 1:200 for the warm front. It should also be noted that these pictures of fronts are idealised and that real warm and cold fronts can have a much more complicated vertical structure.

More complicated vertical structures also occur in occluded fronts, which are common over the UK (Fig. 3). Determining the temperature structure of the air mass and hence the precipitation and cloud structures requires examination and comparison of radiosonde ascents, surface analyses and satellite images.

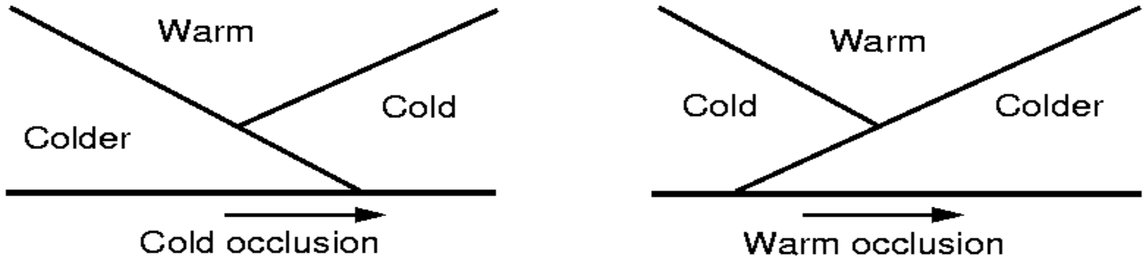


Figure 3: Cross-sections through warm and cold occluded fronts

3.5 Lifecycle of an extra-tropical cyclone

Typical stages in the lifecycle of an extra-tropical cyclone are shown in Fig. 4. Again these stages are highly idealised and each individual cyclone will be subtly different to the others. Extra-tropical cyclones form on the temperature gradient between the equator and pole. In an idealised sense, this gradient can be thought of a one long frontal feature between warm air to the south and cold air to the north (Fig. 4a). Cyclones often form as “so-called” waves which initially look like small kinks in the front but which grow into angular points marking where the cold and warm fronts meet (Fig. 4b). As the system begins to mature, the central pressure at the point where the fronts meet decreases, forming a “warm-sector” with generally overcast cloud conditions but little precipitation (Fig. 4c). During the final stage of cyclone development the cold front begins to catch the warm front at the surface and form an occluded front, close to the centre of the cyclone (Fig. 4d).

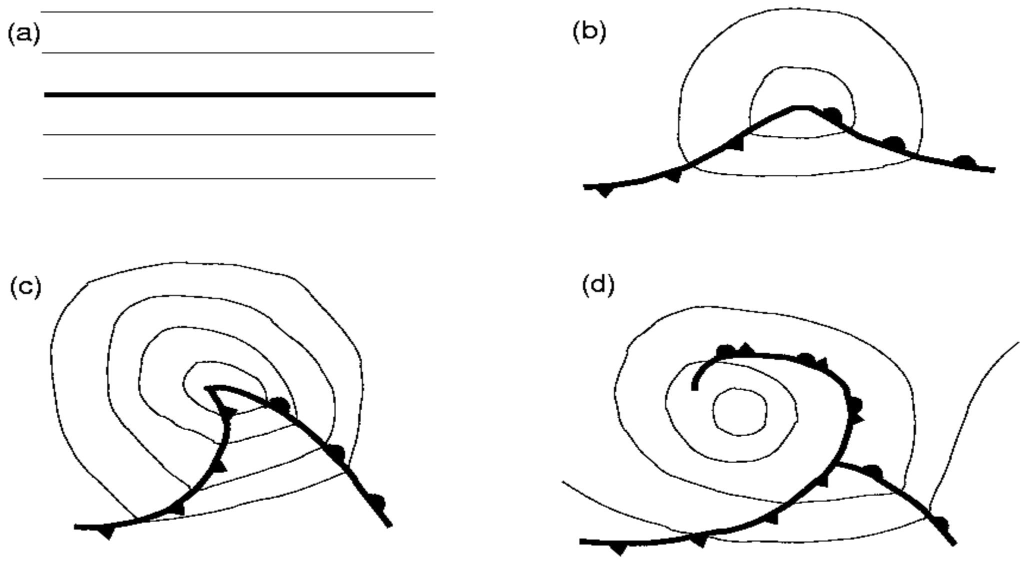


Figure 4: Stages in the development of a mid-latitude cyclone. (a) Pre-existing front. (b) wave forms on front. (c) fully-developed cyclone with 'warm sector'. (d) occluded stage.

So-called secondary cyclones can develop in a similar way on pre-existing trailing fronts of mother cyclones. Their evolution is usually much faster and they are smaller in scale, which makes them a difficult feature to forecast.

4. Local influences on forecasting

4.1 Orographic wind effects

There are many ways in which the presence of steep orography can have significant impacts on the local flow, either through a direct dynamical impact on the flow or through a thermodynamic generation of flow in mountain valleys through different heating and cooling of different parts of the valley. You should be aware of these potential local impacts on your forecast given the location of our measuring site. In particular channelling of the wind down the valley may have important impacts on your wind direction forecasts.

4.2 Orographic enhancement of precipitation

It is often observed that hills as low as 50m can produce significant enhancement of precipitation on the order of 10-40% greater than that upstream of the hill. Although the interaction of orography and the synoptic and mesoscale flow is complex (see Fig. 5 for potential interaction mechanisms), over the relatively shallow hills of Arran, the dominant interaction is the seeder-feeder mechanism. As air is forced to ascend as it meets the hill, a small non-precipitating cloud called a feeder cloud is formed. When a significant precipitating cloud, like that seen in a frontal feature, passes overhead falling precipitation from the seeder cloud can collect more moisture from the feeder cloud before it hits the ground. This results in higher rainfall totals for the same storm in hilly regions.

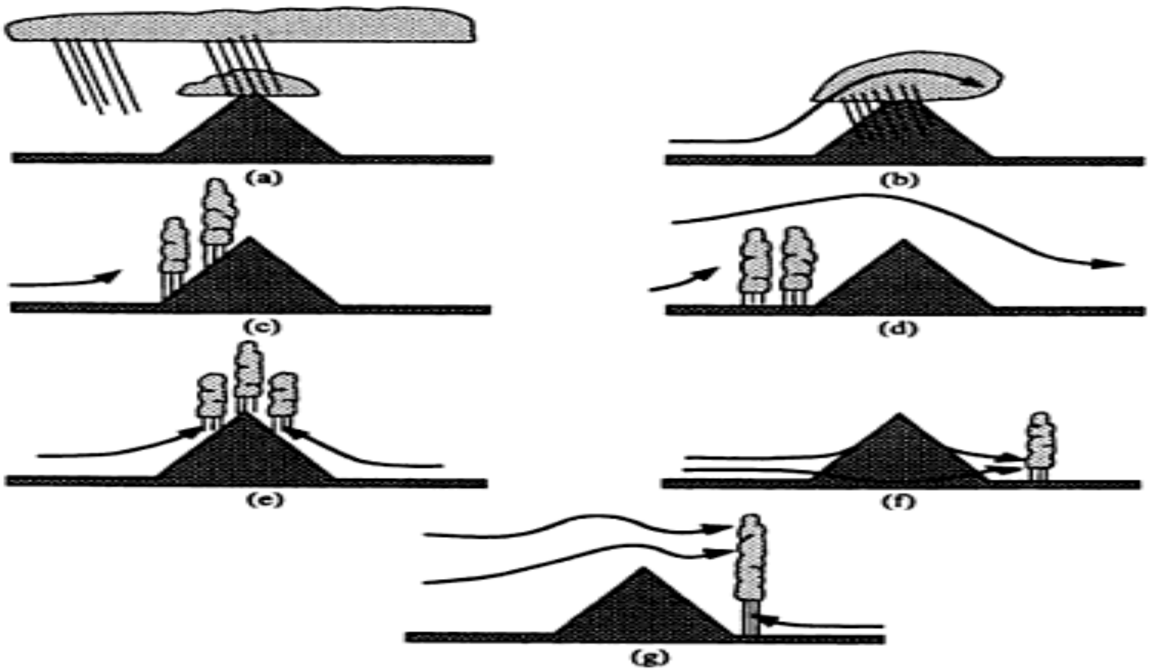


Figure 5 Interaction of orography with precipitation. (a) Seeder-feeder mechanism; (b) upslope condensation; (c) upslope triggering of convection; (d) upstream triggering of convection; (e) thermal triggering of convection; (f) lee-side triggering of convection; (g) lee-side enhancement of convection; Figure taken from Houze, Cloud Dynamics (1994).

5. Available data sources

5.1 Morning weather briefing

We will start each day with a weather briefing at 07:30am. One or more of the staff will present the latest weather charts and discuss how the synoptic situation is developing. Attendance at this is both essential and compulsory. You are encouraged to take an active part in the discussions at the weather briefings. Typically in a real forecasting situation a similar briefing takes place at the change-over of shift between two forecasters or during flight planning in a measurement campaign. You should use the briefing to find out how successful your overnight predictions were and how the synoptic situation has changed overnight.

5.2 Current analysis charts

You will be provided with 6-hourly current surface analysis charts for the previous 24 hours. These should be used to identify the key features of the current state and to determine the direction that these features are moving, and if they are growing or decaying. An example of a current analysis chart is shown in Fig. 1.

5.3 Previous local mast data

Apart from the first day of forecasting there will be data available from the meteorological mast for the time in which we are staying at Lochranza. This data is useful to get an idea of what impact local effects are having on the conditions at the field centre. Comparison of predicted temperatures, windspeeds and rainfall amounts from the large-scale model with the actual observed values gives a rough guide as to the offsets you may need to add to model forecasts to achieve accuracy for our field site. However, some caution must be used when making this comparison as the local effects may also be dependent on large-scale flow features such as wind direction.

5.4 Tephigrams

You will be familiar with tephigram plotting and analysis from your balloon ascent in the course (see the following chapter for more details). There will be 5 ascents from Met Office stations available at 12 hourly intervals for the UK. You should use the tephigrams to examine the position of frontal features over the UK and to determine the details of the air mass dominating conditions in Arran. The conceptual models of frontal features discussed in the previous section will help you to understand the structure of the tephigrams.

5.5 Satellite data

Two different types of satellite images are available, both from the geostationary Meteosat satellite. Geostationary satellites are located at a fixed point above the earth at the equator at an orbital period equal to the rotation rate of the Earth, always measuring the same part of the globe. You will have access to hourly images from the satellite for a projection over Western Europe in the visible and infrared channels (see Fig. 6).

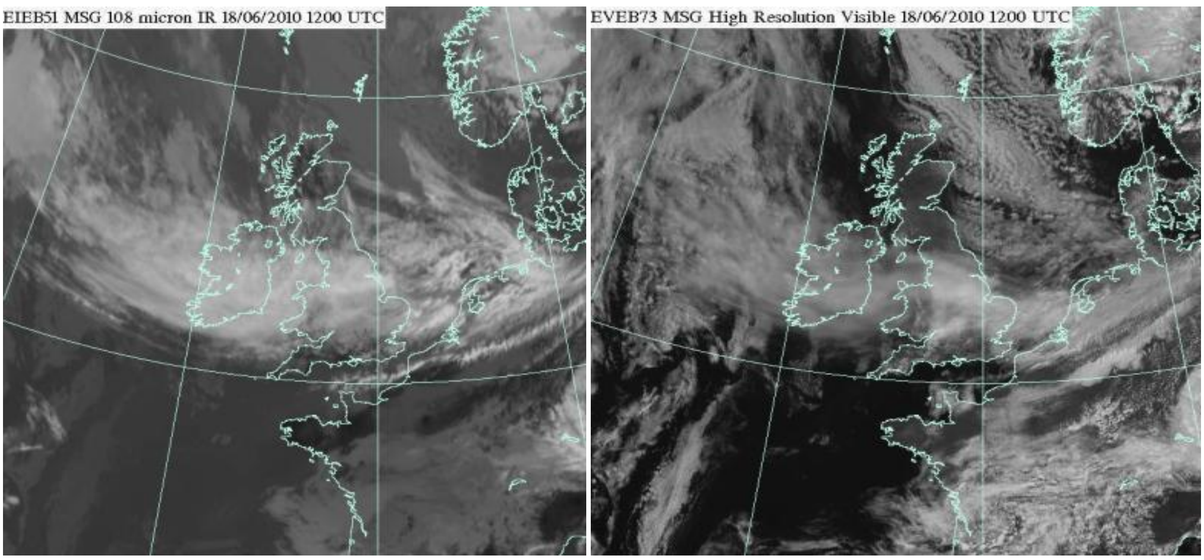


Figure 6: Meteosat satellite images of the UK and surroundings for 1200 UTC 18 June 2010 in the infra-red (left panel) and visible channel (right panel).

- **<u>Infra-red images</u>**

  Infrared images show the invisible infra-red radiation emitted directly by cloud tops and land or ocean surfaces. The warmer an object is, the more intensely it emits radiation, thus allowing us to determine its temperature. These intensities can be converted into greyscale tones, with cooler temperatures showing as lighter tones and warmer as darker. Lighter areas of cloud show where the cloud tops are cooler (and hence higher) and therefore where weather features like fronts and shower clouds are. The advantage of infra-red images is that they can be recorded 24 hours a day (see frontal zone stretching across the UK in Fig. 6). However, low clouds, having similar temperatures to the underlying surface, are less easily discernable (see clouds over France in Fig. 6). Coastlines and lines of latitude and longitude have been added to the images and they have been altered to northern polar stereographic projection.

- **<u>Visible images</u>**

  Visible images record visible light from the sun reflected back to the satellite by cloud tops and land and sea surfaces. They are equivalent to a black and white photograph from space. They are better able to show low cloud than infrared images (low cloud is more reflective than the underlying land or sea surface; see clouds over France in Fig. 6). However, visible pictures can **<u>only</u>** be made during daylight hours.

5.6 Rainfall radar

Images from rainfall radars are extremely useful indicators of the intensity of precipitation. The radars detect the reflected signal in the microwave frequency band from falling precipitation. A network of radars 150 km apart covers the whole of the UK and Ireland. It should be noted that whilst these images give a good indication of the intensity of precipitation, they do not give accurate quantitative measurements. Composite images for the UK radar network will be available at 30 min intervals for the previous 12 hours. An example is shown in Fig. 7, which demonstrates that the frontal zone evident from Fig. 6 is associated with localized heavy rain over central England.

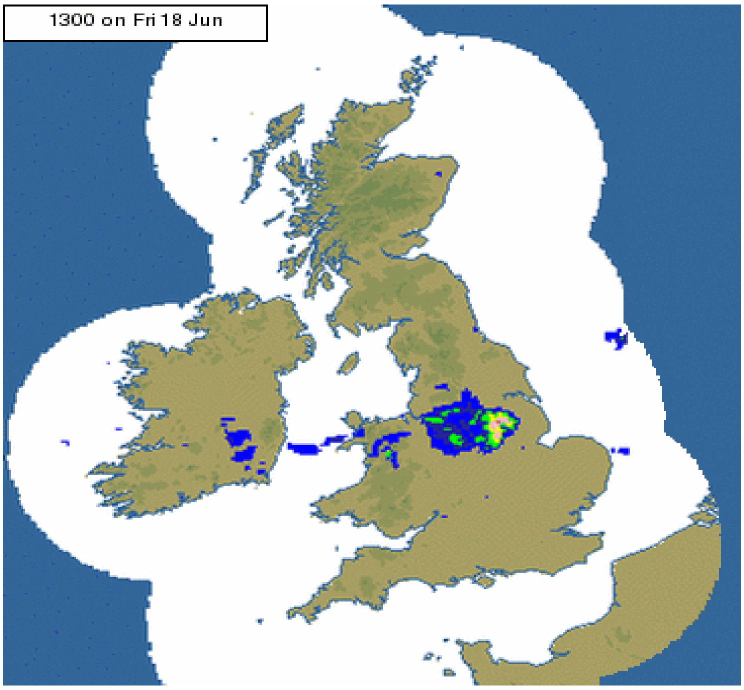


Figure 7: Rainfall radar composite for the UK and surroundings for 1200 UTC (1300 local time).

Warmer colours indicate greater precipitation rates.

5.7 Met Office model forecasts

The main source of forecast information you will have access to is from the Met Office chain of forecast models (see Fig. 8 for an overview). These different models are used to produce forecasts on different timescales and have different strengths and weaknesses. The main difference between the models is in the spacing between points, which are used to represent the flow, and in the area covered by the models. You need to take this into account when you use the model forecasts. It will determine how and what you use and how much confidence you have in the model.

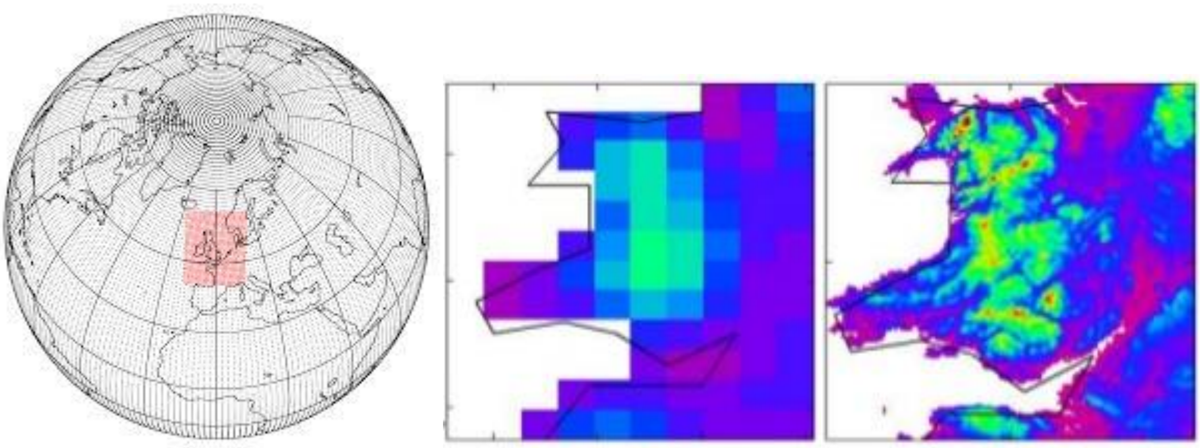


Figure 8: Global and UKV spatial domains (left). Resolution of topography used in Met Office global model (middle) and high-resolution UKV model (right).

- Global model
  The Met Office global model is used to produce the forecasts of surface pressure and frontal features. It represents the entire globe and has a resolution of 0.141 degrees longitude x 0.094 degrees latitude (~10 km) in the horizontal and has 70 levels in the vertical. It is run for forecasts of 144 hours or 6 days. This model is used to give a broad scale overview of the coming weather situation over the next few days. It has good resolution of synoptic features which have typical scales of 1000 km and of atmospheric fronts.

- UK Model
  The Met Office UK model is used to produce the forecast of small-scale convection. It represents an area over the UK and surrounding sea and has a resolution of 0.0135 degrees longitude x 0.0135 degrees latitude (~1.5 km) in the horizontal and has 70 levels in the vertical. It is run for 54 hours because of the extra computational resources required to run at the much higher horizontal resolution. Input data come from the coarser global model. This model does not need a simplified representation of deep moist convection as the global (a so called ‘parametrisation’), since it can explicitly resolve most types of convective showers and thunderstorms.

5.8 ECMWF medium-range ensemble forecasts

On timescales greater than a few days the atmosphere can evolve in a less predictable way than on the short timescales on which you will mainly be making forecasts. Part of the problem in making forecasts on these medium-range timescales is that small errors in the initial analysis used to make forecasts can grow rapidly and non-linearly and change the overall large-scale structure of the flow. One way to account for this uncertainty is to make a large number of forecasts, all with slightly different initial conditions and examine the probability of different forecast outcomes. On Arran we will have available ensemble forecasts for the local area in the form of meteograms generated by the ECMWF Ensemble Prediction System (Fig. 9). This can be used to examine any major changes to the large-scale flow, for example a breakdown of a blocking anticyclone over the UK, which might indicate a change in weather regime during the next few days. The spread of the 50 different ensemble members indicates the reliability or uncertainty of the forecasts provided. The magnitude of the uncertainty depends to a large degree on the prevailing flow situation. You can take this uncertainty into account in the interpretation of the model forecasts.

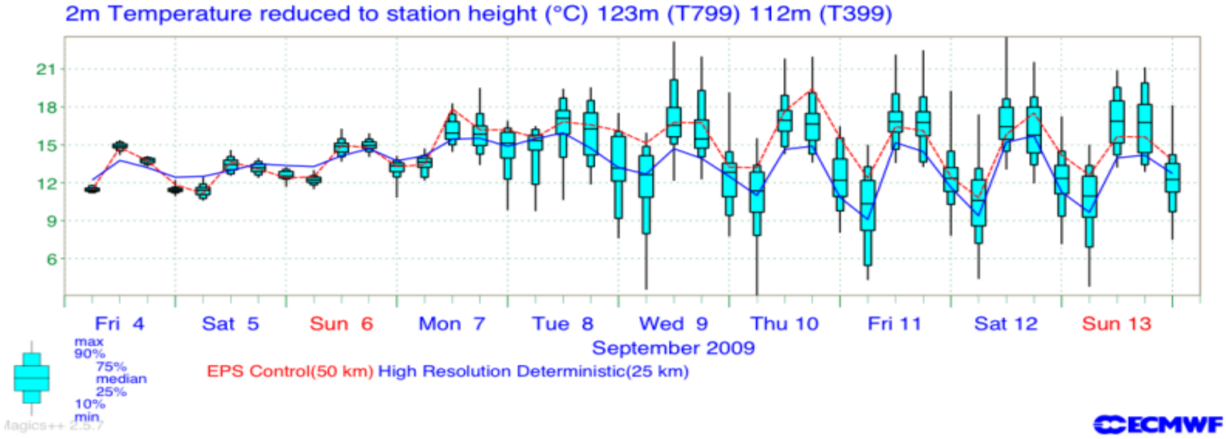
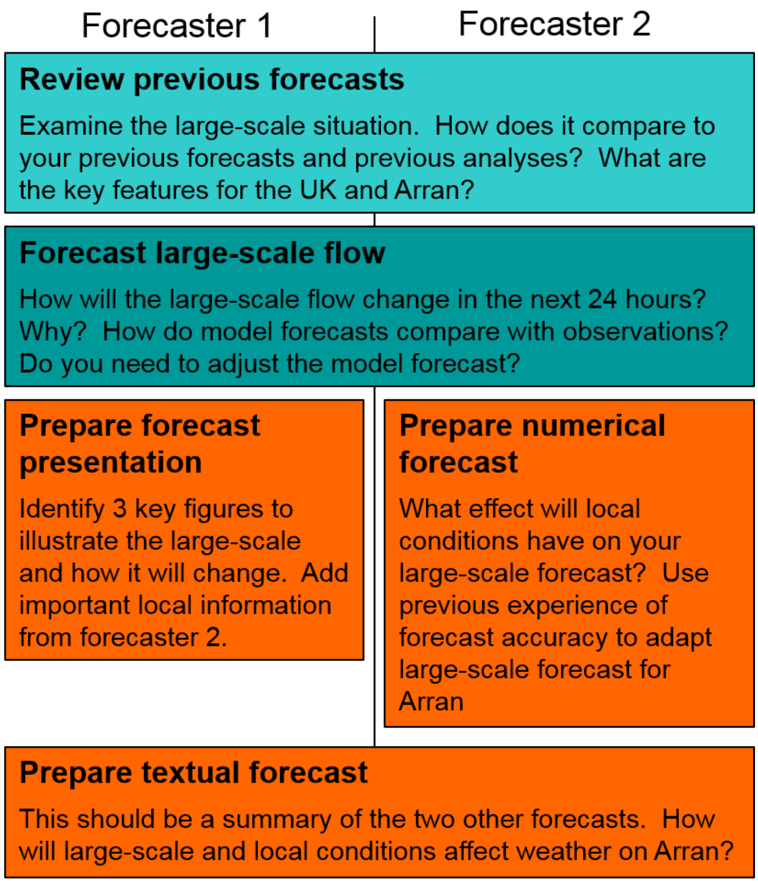


Figure 9: European Centre for Medium-Range Weather Forecasts (ECMWF) ensemble prediction system (EPS) meteogram for Arran for the period 04–13 September (during the field course 2009).

Shown is the 2-m temperature in °C. The blue line is the high-resolution deterministic forecast; the red line the coarser-resolution control from the EPS (i.e. the forecast started from the same initial conditions as the deterministic forecast). The differences between the two are indication of the influence of horizontal resolution on the forecast. The boxes and whiskers show extremes and the 90%-, 75%-, 50%-, 25%- and 10%- percentile of the EPS. The size of the boxes and whiskers (the ‘spread’) is an indication of the reliability of the forecast (high or low predictability of the situation). Note that the time axis is in 6-hour steps in UTC (so these charts don’t necessarily show min/max temperature unless this occurs at 00, 06, 12, etc).

6. Forecasting Practice

This section should be used as a guide for the preparation of forecasts each evening. It is important to develop a procedure for your team so that all of the members know what to do each evening and that your forecasts are consistent. However, different forecasting situations require different approaches and you may find that you need to change the order that you approach the forecast according to the situation.



6.1 Nowcasting of rainfall from radar images

Forecasting the amount of rainfall is one of the more difficult forecast tasks. You will need to use a combination of the rainfall accumulation forecasts from the Met Office model and the radar rainfall (see Fig. 7 for an example). The model is most useful on long timescales (beyond 6 hours). The radar rainfall is useful for short-term forecasts (less than 3 hours). A simple nowcasting method is to assume that any rainfall features currently shown on the radar will move linearly and rain at constant rate over the next 3 hours. By calculating the velocity of rainfall features, their distance from Arran and their spatial scale an estimate can be made of the duration of rain over Arran. Typical rain rates for frontal features in the UK are between 0.5 and 4.0 mm h^{-1} and for non-frontal (showery) features between 2 and 50 mm hr^{-1}

6.2 Forecasting maximum/minimum temperatures, wind direction and speed

Forecasting max/min temperatures relies on accurate predictions of several variables including cloud cover, cloud type, mean wind speed, dew-point temperature and atmospheric stability. In addition, local orographic and land-use effects (e.g. frost/fog hollows, wind channelling, urban heat islands) are not fully accounted for, even in high-resolution forecasts. You will need to consider the accuracy of model forecasts for the previous day(s) and their limitations when preparing your forecasts.

Radiosondes

Radiosondes and tephigram

MARKS

- 100% for analysis

GROUP AND INDIVIDUAL WORK REQUIREMENTS

- You must carry out the analysis of the data and answer the questions yourself as an individual.

- You must write up the answers to the questions in your own words.

Learning objectives

After completing this exercise you will be able to:

a) Understand how to plot and interpret a tephigram.
b) Relate observed sonde data to those from different parts of the UK.
c) Plan a radiosonde launch to address specific scientific questions.

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.

The principal axes of a tephigram are temperature and potential temperature; these are straight and perpendicular to each other, but rotated 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 downwards towards bottom left, so that altitude increases from bottom to top on the diagram. This 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 non-linear variation of saturated vapour pressure with temperature. The saturated adiabats usually extend on to the -40°C level; at temperature 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 Celsius, the potential and equivalent potential temperatures should strictly be in Kelvin, and all calculations involving them should use values in Kelvin.

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

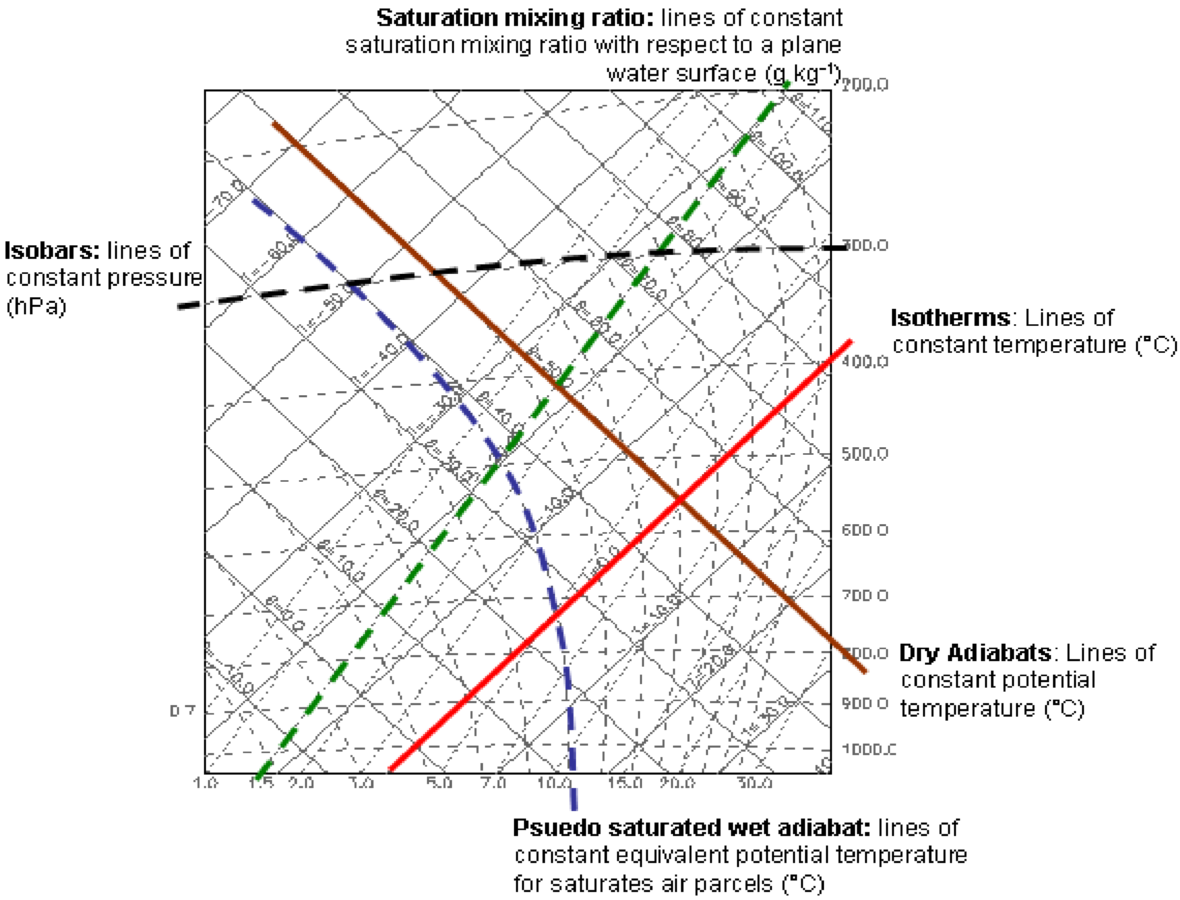


Figure 1. The tephigram, with the principal quantities indicated.

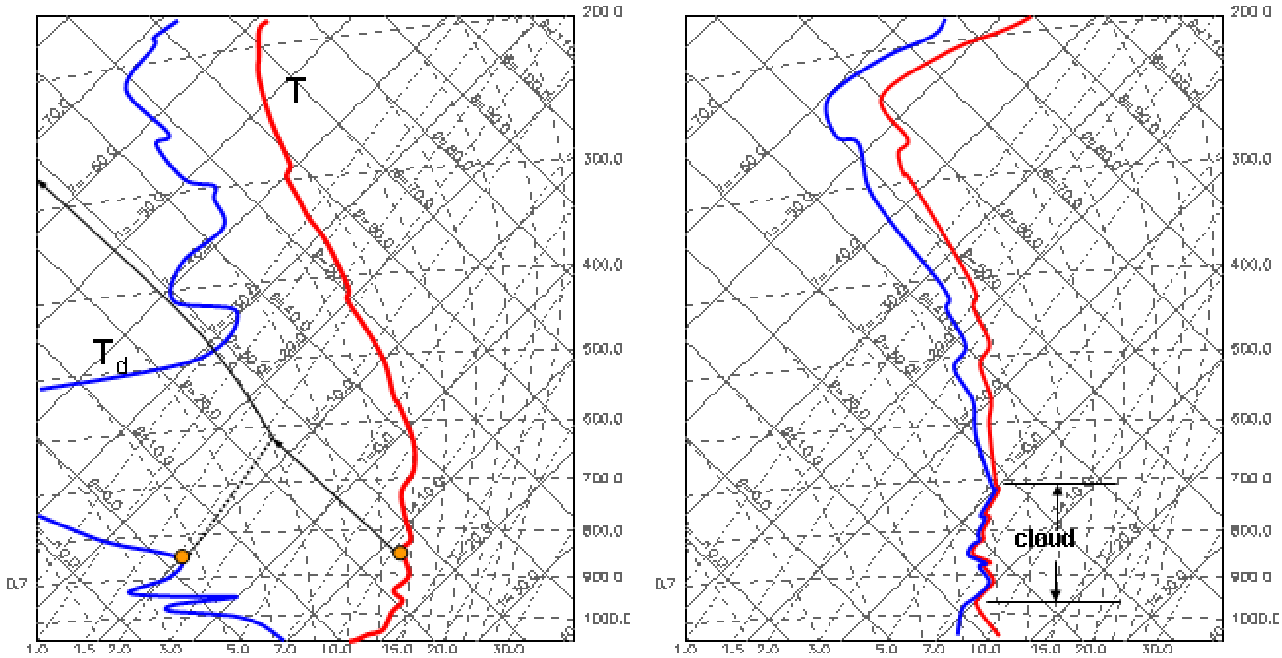


Figure 2. Example tephigrams

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 was 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 could condense out to form a cloud; this is knows and 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 the 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 change 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 adiabatic temperature paths on the tephigraph. A parcel of air that is warmer than the air around it is less dense and therefore buoyant, and the layer is 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 idealized examples below illustrate several different classes of stability.

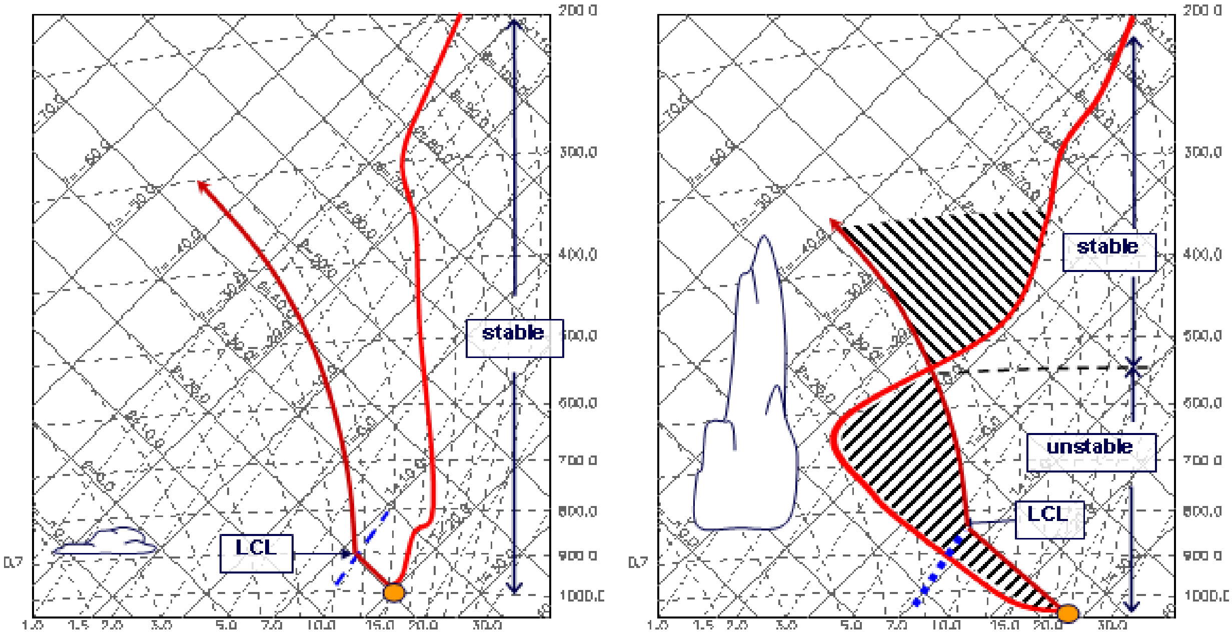


Figure 3. Idealised examples of absolute static stability (a) and instability (b) of air at the surface

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 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 at 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 it original level a number of times, slowly losing energy. Such oscillations are called gravity waves, and often occur where stable air is forced up over orography. If clouds form 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 new-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 hPa; 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 surrounding 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 point during the ascent; but it remains a useful guide to the approximate maximum height of updraughts.

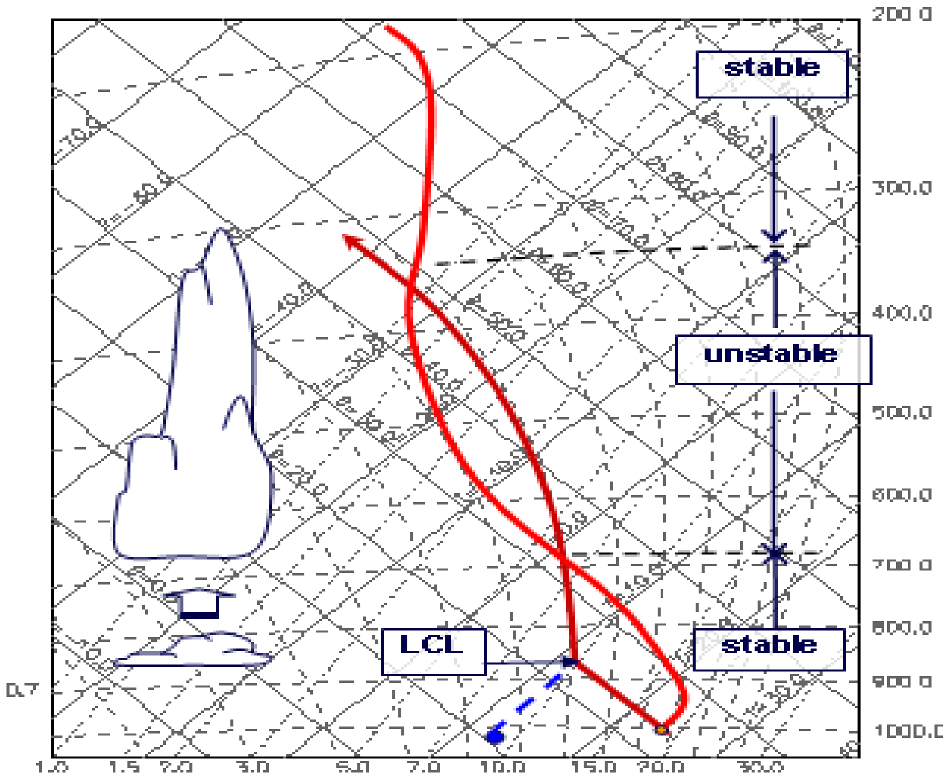


Figure 4. Idealised example of conditional instability

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 remains dry, it would remain stable no matter how high it was lifted; however, in this case the LCL is reached at around 850 hPa, and the rate of cooling decreases to that of the saturated adiabat. The saturated adiabat intersects the environmental temperature profile at about 680 hPa; 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 340 hPa 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 of stability.

Exercise

1. <u>Planning a radiosonde launch (35% of your final mark)</u>

A key part of any fieldwork activity is planning when and where to deploy your expensive and sensitive equipment. Early on in the week, in this part of the activity, you will write a short proposal for when to deploy an additional radiosonde, taking into account the restrictions on your launch below.

Each group will have one hour to design a short project proposal for when to launch the radiosonde. Each group will produce a proposal of 1 side of A4 which should be given to the activity leader by (**TIME TO BE ANNOUNCED**). The proposal should contain the following information:

- When you would like to launch the radiosonde?
- What scientific questions you hope to answer?
- The reasons behind your launch choice i.e. why you think you will be able to answer the scientific question you propose.
- Is there any uncertainty in your proposal? What have you done to minimize the uncertainty.

When designing you experiment, a critical part of your choice will be the meteorological conditions predicted for the day in question. You can make use of the forecast charts provided for the weather forecasting activity and the other data sources described later. Some of the things that you might choose to look at with your radiosonde are:

1. What are the characteristics of the airmass over Arran? Is it warm or cold, moist or dry? Is it unstable or stable?
2. What is the position of a predicted or analysed frontal feature? How close is it to Arran?
3. Where are cloud layers located over Arran? Can additional properties of the cloud (for example how much precipitable water) relevant for us at the surface be derived from the ascent?
4. Can the radiosonde be used to test the accuracy of a model forecast at a critical time for other activities?

RESTRICTIONS ON THE LAUNCH ARE AS FOLLOWS:

Once each case has been submitted, a steering committee made up of members of staff on the course will decide which makes the strongest scientific case and award the additional radiosonde launch to the winning team. Once the decision has been made, the chair of the steering committee will make a short (10 min) presentation to the group outlining the reasons behind their decision. A document outlining the decision will be placed on the shared drive.

The winning team will then launch their radiosonde and the results of the ascent will be plotted on a tephigram and available to the group soon after the experiment is completed.

Once the experiment has been completed you should write up this part of the activity in your notebook (no more than 2 pages), including the following parts:

a. Your proposal
b. A comparison of your proposal with the winning proposal. Include why you think the winning proposal was chosen, if you agree with the decision and why or why not?
c. The resulting ascent
d. If the experiment allowed us to answer the scientific question posed. If it did, what was the answer to the question posed. If it didn't, why did the experiment fail?

2. Plotting and analysing a tephigram from the Saturday radiosonde launch (25% of final mark)

You will have access to a range of radiosonde launches from Lochranza and from other UK sites throughout the week which have been automatically plotted on tephigrams for you. To help your understanding of what is plotted and how to analyse a tephigram, we will ask you to manually plot the profile for the radiosonde we launch on Saturday. The data from this ascent will be made available on the shared drive. Ask the activity leader for a blank tephigram and plot the ascent. Once you have plotted the tephigram, please complete the following analysis of the situation.

a. Write a brief description of your ascent identifying features of interest e.g. cloud layers, frontal surfaces etc. Make sure you also consider the surface analysis chart for the same time.

b. Estimate the lifting condensation level for near-surface air. Note that radiosonde measurements right at the surface frequently display some errors due handling of the sonde by the operator during the launch.

c. Determine whether conditions are stable or unstable. If conditions are unstable, estimate the level to which convection is likely to extend. Are convective clouds likely to form?

3. Your own radiosonde launch (40% of your final mark)

In addition to your own sounding at Lochranza there will be other soundings available from around the UK taken at approximately the same time (see figure 5).

a. Starting with the 0000 UTC surface analysis and the satellite imagery provided, describe briefly the synoptic situation across the British Isles at this time. Highlight which parts of the British Isles are affected by different air masses (if any). (Write less than ½ page).

b. Calculate the wet-bulb temperatures at a selection of heights from your Lochranza sounding, as well as two other sounding stations, and provide a table. Based on your table, identify two sounding stations for which you expect air masses to have different origins. For these locations, specify the heights for which you would like to study backward trajectories. (Write less than ½ page).

Backward trajectories are provided and can be accessed online. Include your chosen trajectories in your workbook.

Figure 5: UK Radiosonde launch sites. Hers.=Herstmonceux.

c. Choose one synoptic chart from the 5 days leading up to your sounding and include it in your workbook. Using the chart and your trajectories, describe the origin and evolution of the air masses that you identified in (a). In particular, you should note if and when the trajectories diverge, and if trajectories follow any particular ascending or descending motion, then relate these changes to the synoptic situation, e.g. particular locations of fronts and low pressure regions. (Write less than 1 page).

Hill Profiling

Hill profiling

MARKS

- 30% for performance in field
- 70% for analysis

GROUP AND INDIVIDUAL WORK REQUIREMENTS

- You may share raw data from the instruments and field notes with the whole group.
- You may share tables/plots of data with the whole group.
- You must carry out the analysis of the data and answer the questions yourself as an individual. You must write up the answers to the questions in your own words.

Learning Outcomes

At the end of the exercises in this section, you will be able to:

- plan an observational programme and respond to changing local conditions.
- describe the stability properties of the boundary layer over Arran
- estimate how Goatfell influences the winds over Arran.
- test the suitability of both the dry and moist hydrostatic approximation in explaining the vertical variation of temperature.
- understand the influence of wind speed on barometric altitude derivation
- communicate effectively with rest of team in the field.

Aim of the activity

In most weather systems, variation in the horizontal direction is much slower than that in the vertical, where rapid fluctuations in atmospheric properties are observed as we ascend. For this reason, balloon ascents through the atmosphere (up to about 25 km) are made routinely around the globe. Unfortunately, although this data is of great importance to forecasters, the cost of the operation limits radiosonde ascents to about 12 stations in the UK (see map on the computer system). As part of our field exercise we will be measuring the atmospheric profile using both radiosondes and/or by ascending some steep hills around the field centre.

What you will do

We can start to answer these questions above by making measurements using hand-held meteorological instruments as your groups ascend Goat fell. You will have the following:

- Digital barometer
- Whirling psychrometer (dry-bulb & wet-bulb thermometer)
- Digital thermometer
- Cup anemometer
- Compass/OS map
- GPS receiver

During Saturday you will have chance in your groups to learn how to use these instruments. On the ascent you will need to measure the following key quantities:

- Time (GMT)
- OS position (using the GPS receiver and/or map) in Northings and Eastings
- GPS Altitude
- Pressure
- Temperature (dry & wet bulb)
- Wind Speed/Direction
- Prevailing weather conditions

Before you leave for the ascent of Goatfell you will then need to design a sampling strategy for your group. The sampling strategy should cover:

A. Where and how frequently to take measurements? You should think about how many measurements will be needed to profile the atmosphere on Goatfell and where good measurement sites are located (based on the map) . You should also consider if you should take measurements on the ascent only or also on the descent and why.
B. How you will measure the key quantities above? You should think about which instruments will you use for each quantity, how uncertain each measurement is and how many repeat observations are needed. You should also develop a team strategy for how measurements will be taken and who will do what.

Once agreed and discussed with your team leader you should write a short sampling strategy (around one page) in your handbook. Be careful to include your reasoning for the decisions made. **The sampling strategy you use will be the first part of the assessment for this activity**

On returning to the field centre you will need to process your data to come up with a final spreadsheet of your observations. Please complete this as soon as possible and add your measurements to the shared spreadsheet *hill_data.xlsx* under the appropriate group name. The spreadsheet should also calculate some moisture variables based on your inputs in the tab *group name calcs*, check that these values are reasonable before saving the sheet.

Once you have completed this task move on to the exercises at the end of the section. You will need to have read/consult the background information below as you complete the exercises.

Some important information about altimeters

You will have an altimeter (height derived from ambient pressure measurement) available to find your approximate altitude on Goatfell. However, you cannot use this to obtain height data for plotting in this exercise. The reason is that the altimeter has the hydrostatic equation programmed into it already, so you will not obtain an independent verification of the equation. Instead, you must use GPS for the altitude.
For verification of the hydrostatic equation, you can use the GPS altitudes directly. These are generally accurate to about +/- 20 m. However, for the accurate measurement of the height of Goatfell, this is not good enough. We need an accuracy of significantly better than 7 m, since we can measure the pressure to 1 hPa and the pressure changes by about 1 hPa in 7 m altitude.

Before setting out from Lochranza, you should set the altimeter using the known height of the centre (road outside is at about 5 m AMSL). *Why is it necessary to set an altimeter like this?*

At the same time as setting your altimeter, make a note of the pressure reading of your altimeter and also of the pressure reading on the precision barometer in the classroom. Repeat this when you return. You will need this data to correct for any synoptic pressure changes which may occur during the walk. *Has there been a change in pressure throughout the day? Why is this?*

Background information

The boundary layer

The lower part of the atmosphere in which most weather systems occur is called the *troposphere.* Above this lies the *stratosphere*, and the boundary between these layers is the *tropopause*. The tropopause height usually varies between 8 and 12 km but can descend almost to the surface in cold frontal events and reach above 17 km over the tropical continents. The lowest part of the troposphere is the *boundary layer*, which is loosely defined as the part of the atmosphere that is directly influenced by the surface. The boundary layer depth may vary from tens of metres to a few kilometres above the surface. The boundary layer depth is influenced by diurnal heating patterns and complex terrain like the hills of Arran, and may thicken or contract over a mountain depending on the wind speed and stability of the atmosphere. Note that because we will never lose contact with the surface during our walk up Goatfell, we will likely remain within the boundary layer the entire time.

Potential Temperature and stability

Air pressure decreases with height and temperature generally also decreases. Even if the airflow is very turbulent and thoroughly mixed vertically, this is still true. The reason is that if a parcel of air is moved upwards it expands, and sudden expansion without addition of extra heat results in a lowering of the temperature (by the ideal gas law). Potential temperature (denoted by θ) is a quantity which does not change for an air parcel unless we add or remove heat (e.g. by solar heating). It is related to the pressure p, the absolute temperature T (in Kelvin) and the surface pressure p_s by

$$\theta = T\left(\frac{p}{p_0}\right)^{-0.288} . \qquad (1)$$

Physically, θ is the temperature an air parcel would have if it were brought down or up to a level where the pressure is p_0=1000 hPa (in practice, close to the surface) and hence compressed or expanded without adding heat. Because potential temperature doesn't change for an air parcel unless heat is added to it, we call it a *conserved quantity*.

Within the boundary layer, strong turbulent mixing due to surface heating and/or frictional shear stress often gives rise to a nearly constant vertical θ profile, whereas above the boundary layer θ generally increases with height. By examining the "environmental" vertical profile of θ (such as that measured by a radiosonde) we can determine the *static stability* of the atmosphere. When the environmental θ (or θ_{env}) increases with height, air parcels that are vertically displaced within the layer will tend to return to their initial position, which is a *statically stable* situation. Consider a small volume, or "parcel" of air within a stable layer that is displaced upwards (left panel of Fig. 1 below). The potential temperature of the parcel (θ_p) is initially equal to that of its environment, and remains unchanged as the parcel is displaced because its θ is conserved. Because θ_{env} increases with height, the lifted parcel becomes colder than its environment ($\theta_p < \theta_{env}$). As a consequence of the ideal gas law, this relatively cold parcel is also denser and heavier than its environment and hence *negatively buoyant*. As shown by the blue arrow, this negative buoyancy ($\vec{b}$) returns the parcel back towards its original position. On the other hand, if the parcel is displaced downwards from its initial position, it becomes positively buoyant and is drawn upwards back to its original position. Thus in a statically stable situation a parcel of air tends to stay where it is and if it is moved it tends to return to its original position (in the vertical).

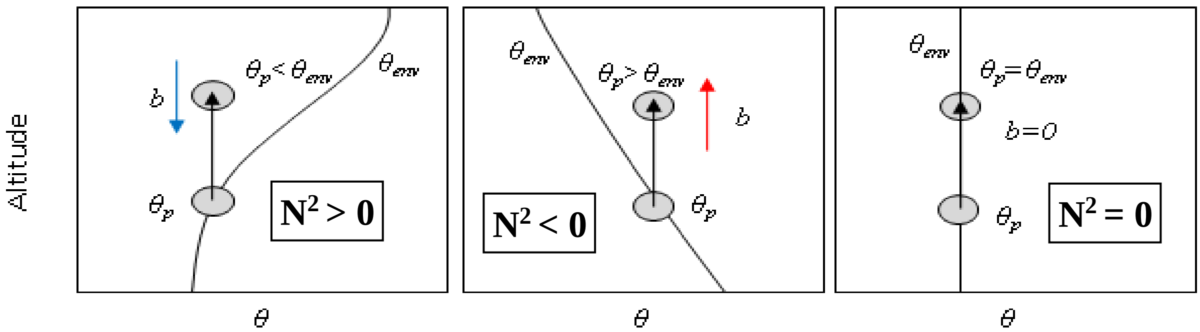


Figure 1: Vertical profiles of potential temperature (θ) and illustration of parcel stability

By contrast, when θ_{env} decreases with height, an air parcel that is displaced upwards (downwards) will become positively (negatively) buoyant and will continue rising (sinking). This is a *statically unstable* situation because the parcels continue to accelerate in the same direction as their displacement. When statically unstable layers develop in the atmosphere, such as when the ground is heated by the sun during the daytime, convective eddies develop that rapidly mix θ within the layer to neutralize this instability.

A *statically neutral* layer is one where θ_{env} remains constant with height. Air parcels that are displaced vertically through a neutral layer will remain at their new level without undergoing any buoyancy-driven acceleration because their θ remains identical their environment. Due to turbulent air motions that vertically mix potential temperature (and other properties), the boundary layer is often characterized by neutral stability.

Note that the *above analysis of parcel stability only holds when the parcel is unsaturated* (ie there is no liquid water in the air parcel). If the parcel saturates (the relative humidity reaches 100%) due to lifting (say, as it ascends a mountain), latent heat is released within it due to condensation and/or deposition that causes its θ to increase. Likewise, a parcel's potential temperature decreases if it experiences evaporation and/or sublimation of liquid or frozen water particles.

You will be able to plot potential temperature against height from both the hill walking exercises and from the radiosondes.

The hydrostatic equation

To a good approximation, the atmospheric pressure varies with height according to the hydrostatic equation. This essentially states that the pressure at any height balances the weight of overlying atmosphere and so neglects the pressure fluctuations associated with vertical air motion.

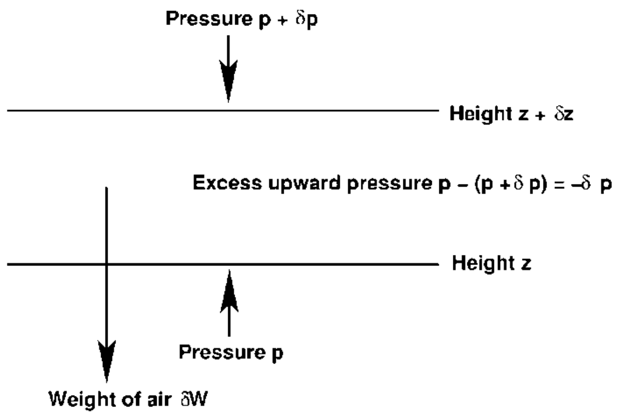


Figure 2: The balance of forces that gives rise to the hydrostatic equation

Because pressure decreases with height, the upward pressure on the bottom of a thin layer of air is slightly greater than the downward pressure on the top. This creates an upward force that, when balanced by the weight of the layer, prevents the air from accelerating vertically. In the figure above, consider a column of the atmosphere of cross-section area A. Between the two height levels z and $z+\delta z$, the weight of the air column is

$$\delta W = \rho g A \delta z \,, \quad (2)$$

and the upward pressure force on the layer is

$$\big(p-(p-\delta p)\big)A = -A\delta p, \quad (3)$$

where p is the pressure, ρ is the density, and g is the acceleration due to gravity (9.81 m s^{-2}). For the hydrostatic balance to hold, these two forces must balance each other, i.e.

$$\rho g A\delta z = -A\delta p. \quad (4)$$

Dividing through by $A\delta z$ and taking the limit as $\delta z \rightarrow 0$ gives the familiar hydrostatic equation

$$\frac{\partial p}{\partial z} = -\rho g. \quad (5)$$

The hydrostatic equation is often combined with the ideal gas law, which is written as

$$p = \rho RT, \quad (6)$$

where R depends on the chemical composition of the air. As the molecular mass of the gas increases, R decreases. This is meteorologically important because moist air contains water vapour (H_2O), which has a lower molecular mass than the primary molecules in air (N_2 and O_2) and hence a larger R. At a given p and T, (6) shows that moist air will be lighter than dry air. For completely dry air R is given by R_d = 287 J kg^{-1} K^{-1}, while for pure water vapour R_v = 461.5 J kg^{-1} K^{-1}.
[NB this is the meteorologist's version of the ideal gas law. However as all good chemists know the ideal gas law is really pV=nRT (where p is the pressure of the gas (Pa), V is the volume of the gas (m^3), n is the number of moles of gas, R is a universal constant (8.314 J K^{-1} mol^{-1}) and T is the temperature (K)) where R doesn't depend upon the chemical composition. But if you multiply both sides of this equation by the molecular mass of the air you get pVM=nMRT, dividing this by V and redefining R as R/M gives you p=ρRT the meteorologists ideal gas law. Unfortunately in this formulation R isn't really a constant as it varies with the composition (mainly water vapour) of the air]

When applying (6) to moist air the convention is to use R_d and to replace T with a fictitious temperature called the *virtual temperature* T_v. This represents the temperature to which dry air would have to be heated to have the same density as the moist air being considered (as warm air is lighter than cold air). T_v can be estimated from the following expression:

$$T_v \approx T(1 + 0.61q_v), \quad (7)$$

In (7) q_v is the *water-vapour mass mixing ratio* and represents the ratio of the mass of water vapour to the mass of dry air in the parcel. *q_v is conserved for unsaturated air motions*. This does not hold for saturated air motions because water vapour may be gained or lost as water changes phase between gas, liquid, and solid form. For a moist atmosphere the ideal gas law can be rewritten using T_v and can then be substituted into (5) to give the hydrostatic equation:

$$\frac{\partial p}{\partial z} = -\frac{pg}{R_d T_v}. \quad (8)$$

If we measure pressure p_1 and height z_1 and pressure p_2 at height z_2, then we can integrate (8) to give

$$\int_{p_1}^{p_2} \frac{dp}{p} = -\int_{z_1}^{z_2} \frac{g}{R_d T_v} dz \ . \quad (9)$$

On the right side, T_v is strongly dependent on height, but the range of absolute temperature variation is small in relative terms (e.g., just a few degrees compared with absolute values of 280-290 K). Hence, we can approximate the variable T_v by a constant, average value $\overline{T_v}$ (take, for example, the virtual temperature halfway up the hill). Evaluating the integral in (11) then gives

$$\ln p\Big|_{p_1}^{p_2} = -\frac{g}{R_d \overline{T_v}} z\Big|_{z_1}^{z_2} \ . \quad (10)$$

Hence

$$\ln\left(\frac{p_2}{p_1}\right) = -\frac{g}{R_d \overline{T_v}}(z_2 - z_1) = -\frac{z_2 - z_1}{H_p} \quad (11)$$

where H_p is often referred to as a "scale height" for pressure or density and is defined as

$$H_p = R_d \overline{T_v} / g \ . \qquad (12)$$

This can be rearranged to give

$$z_2 - z_1 = -H_p \ln\left(\frac{p_2}{p_1}\right) \quad (13)$$

Thus the difference in height between two points in a hydrostatic column of air is proportional to the logarithm of the ratio of their pressures.

Gap flow and bernoulli's equation

Air is commonly observed to travel faster and maintain lower pressure within gaps and valleys in a mountain range than over surrounding areas. These increased wind speeds follow straightforwardly from conservation of mass and momentum. Mass conservation requires that the flux of air entering a control volume must equal the flux of air exiting it. Consider figure 3, which shows air funneling through a gap such as that between two adjacent mountain ridges.

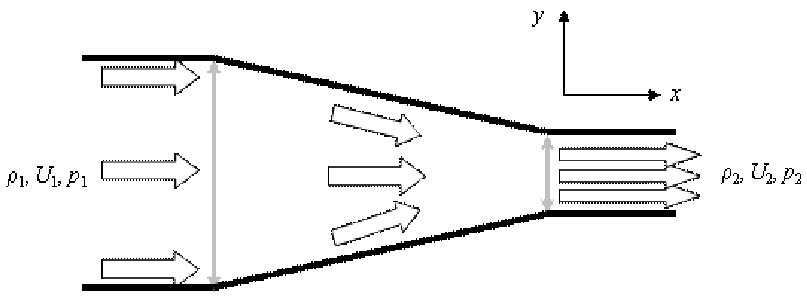


Figure 3. Planform view of horizontal flow acceleration through a gap

On the upstream side, air with a density of ρ_1 and a wind speed of U_1 is confined between two rigid walls separated by a distance L_1. This air must squeeze together as it passes through the narrowing channel and exits the other end of width L_2 with a density of ρ_2 and a speed of U_2. If all of the air displacements are confined to the horizontal plane in the figure, conservation of mass requires that the flux of mass entering (density x velocity x distance) the volume (between the gray lines above) must equal that exiting it, or

$$\rho_1 U_1 L_1 = \rho_2 U_2 L_2 . \qquad (14)$$

If we further assume that the flow is *incompressible*, or that the density of an air parcel does not change along its path, then $\rho_1 = \rho_2$, which gives the following simple relation:

$$U_2 = U_1 \frac{L_1}{L_2} . \qquad (15)$$

Thus the flow accelerates within the gap by an amount that is determined by the width of the gap. If we further assume the flow to be frictionless and steady in time, energy conservation allows *Bernoulli's equation*

$$p + \frac{1}{2}\rho U^2 + \rho g z = \text{constant along a streamline} . \qquad (16)$$

to hold for each parcel of air that passes through the gap (the parcel path is described as a 'streamline'). Because the flow is horizontal and incompressible, the third term on the left side is constant so that

$$p + \frac{1}{2}\rho U^2 = \text{constant along a streamline} . \qquad (17)$$

The above may be rewritten as

$$p_1 + \frac{1}{2}\rho_1 U_1^{\ 2} = p_2 + \frac{1}{2}\rho_2 U_2^{\ 2} . \qquad (18)$$

For incompressible flow this may be rearranged to give

$$p_2 = p_1 + \frac{1}{2}\rho_1 \left(U_1^2 - U_2^{\ 2} \right) , \qquad (19)$$

which may be further simplified by substituting (16):

$$p_2 = p_1 + \frac{1}{2}\rho_1 U_1^2 \left[1 - \left(\frac{L_1}{L_2} \right)^2 \right] . \qquad (20)$$

Because $L_1 > L_2$ the pressure at the gap exit must be lower than that of the impinging flow. Note that this effect (often called the *Bernoulli effect*) explains the changes in wind velocities when air detours around the sides of a mountain peak (Fig. 4). The incoming air, which is unable to pass over the mountain, is thus forced to converge around the sides of the mountain where it passes through a smaller effective volume and accelerates.

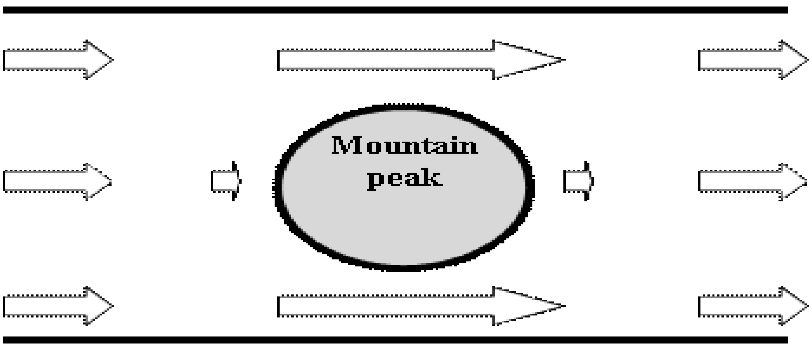


Figure 4: Schematic illustration of flow deflection around a mountain peak and the Bernoulli effect. When winds are forced to detour around a mountain peak, they slow down just upwind and downwind of the peak and speed up around its edges.

Orographic flow regimes

In fluid dynamics and meteorology, we often describe situations in terms of 'dimensionless numbers'. These unit-less numbers, which are derived from the governing equations of motion, describe the dynamical similarities between different flows with different physical scales. One very useful dimensionless number for mountain meteorology is the *Froude number* (*Fr*), which indicates whether a flow impinging on a mountain will rise over it ("unblocked flow") or detour around it ("blocked flow") (see figure 5 below). The Froude number is given by

$$Fr = \frac{U}{NH}, \qquad (21)$$

where U is a characteristic upstream wind speed, H is the mountain height, and N is the Brunt-Väisälä, or 'buoyancy' frequency, which is given by

$$N^2 = -\frac{g}{\rho_0}\frac{d\bar{\rho}}{dz} \approx \frac{g}{\theta_0}\frac{d\bar{\theta}}{dz}, \qquad (22)$$

where $\bar{\rho} = \bar{\rho}(z)$ and $\theta = \bar{\theta}(z)$ are the vertical density and potential temperature profiles of the undisturbed flow and ρ_0 and θ_0 are reference density and potential temperature values (say, that of air at the surface). Note from Fig. 1 that $N^2>0$ when the atmosphere is statically stable (because θ increases with height), $N^2<0$ when the atmosphere is statically unstable, and $N^2=0$ when the atmosphere is statically neutral.

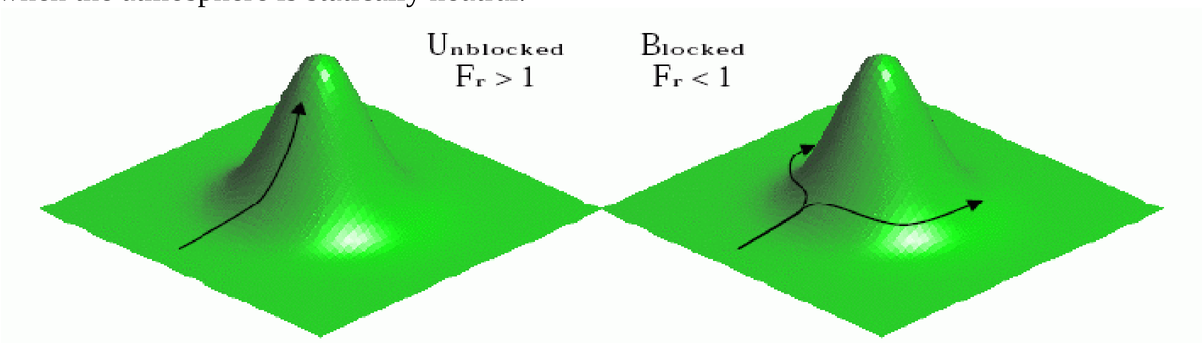


Figure 5: Comparison of low-level responses to an isolated mountain in the unblocked (left) and blocked (right) regimes

The simplest interpretation of *Fr* is that it represents the ratio of the kinetic energy of the impinging flow to the potential energy required to ascend the mountain (i.e., is the airstream moving fast enough to propel itself over the mountain). Because density (θ) tends to decrease (increase) with height, N^2 is usually positive and the atmosphere is *stable*; that is, when air parcels are displaced vertically, they accelerate back towards their original position. When stable air is forced to ascend a mountain, it becomes cold and dense relative to its surroundings (or "negatively buoyant", see Fig. 1), and is pulled downwards.

<u>Unblocked flow</u>
When *Fr* >1 the impinging flow possesses enough momentum to overcome the negative buoyancy that it acquires over the upwind slope, which allows it to complete its ascent over the mountain. In this "unblocked" flow regime, the wind may either accelerate or decelerate on the upwind side of the mountain depending on the value of the *Scorer parameter* (l), which is defined as

$$l^2 = \frac{N^2}{U^2} - \frac{1}{U}\frac{d^2u}{dz^2}, \quad (23)$$

where $u(z)$ is the vertically-varying upstream wind speed (this can be taken from a nearby radiosonde profile), and L is the windward width of the mountain along the main flow direction. According to theory, when $L < 2\pi/l$, the flow accelerates over the upwind slope and decelerates back to its upwind speed over the lee (the opposite is true when $L > 2\pi/l$). The vertical flow structure in each case is illustrated by the diagram below, which for simplicity neglects the second term of (23). When the impinging winds are strong, the stability is weak, and/or the mountain is narrow (left panel), the flow passes over the mountain too rapidly to react to the negative buoyancy that it accumulates over the upwind slope. It responds by squeezing together in the vertical as it passes over the mountain, which causes it to accelerate by the Bernoulli effect discussed above. This response is often referred to as "evanescent" because the mountain disturbance decays rapidly above the ridge. By contrast, when the flow is weaker, the stability is stronger, and/or the mountain is wider, the flow spends enough time over the mountain to react to the negative buoyancy generation. Ascent over the upwind side of the mountain is resisted by the negative buoyancy that is created by the stable lifting ($N^2 > 0$), which causes the flow to decelerate. Over the lee slope the flow accelerates because the negatively buoyant air is free to descend. The "mountain waves" created in this situation are not confined to the area directly over the mountain, but propagate deeply into the atmosphere. Their upward propagation of energy causes them to tilt upstream with height (see Fig. 5). Mountain waves are associated with many important atmospheric phenomena including downslope windstorms, trapped waves, internal wave breaking, and a drag on the atmospheric circulation.

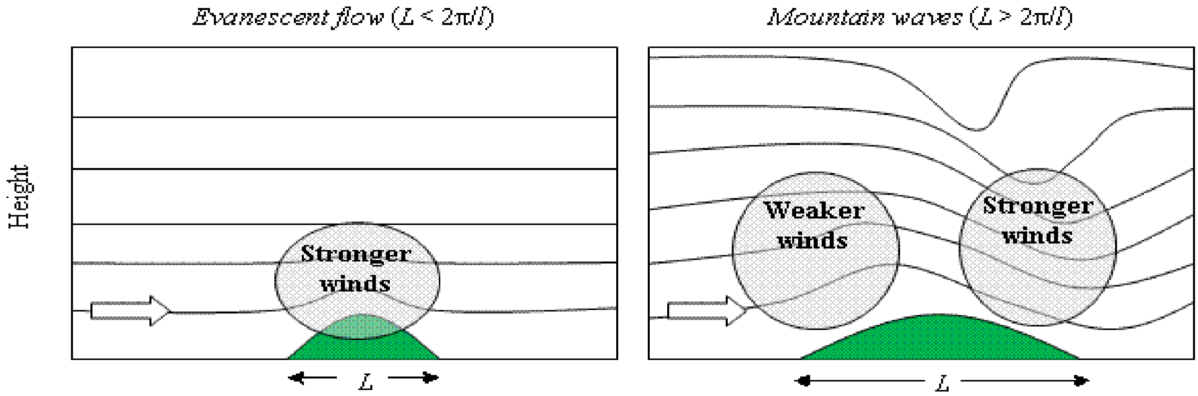


Figure 5: Comparison of the vertical structures of evanescent (left) and mountain-wave (right) flows

<u>Blocked flow</u>
When *Fr* <1, the flow is too weak to overcome the retarding effects of negative buoyancy, so it detours around the mountain. In this "blocked" flow regime, the low-level winds tend to lie parallel to contours of terrain height rather than across them. For this reason, wind speeds are generally low on the upwind side of the mountain but may increase around the mountain edges

due to the Bernoulli effect (see Fig. 4). The winds may also strengthen over the crest due to the ambient increase in wind speed with height and/or wave activity aloft.

Exercises

How to analyse the data

Once all of the data is available you will have a large number of observations from all six groups with which to answer the three questions posed in the introduction. Which of the data should you use? A good starting point would be to attempt to construct profiles using all available data which you think represents a similar background synoptic flow arriving at the hill. You can think of the observations as representing a moving profile through the atmosphere as the different groups ascended. To understand if this is the case you might first compare measurements taken at different times by different groups at similar locations. If in this comparison, conditions change appreciably then you should think about partitioning the data into two different groups representing different synoptic situations. You should also be careful to spot any differences in observational technique between groups. Try plotting profiles of the same quantity from different groups to see if you can spot a consistent offset or bias between datasets. Once you have done some exploratory analysis of the data you should collect all of the data you wish to use into a single spreadsheet tab, ready to do the analysis below. Remember that you will need both the raw and calculated quantities for the analysis.

Before starting this analysis, you should write down which data you chose to use and why (one page only), this will also form part of your assessment [10 marks].

Analysing the stability properties of the boundary layer over Arran

1. For your chosen observations plot the potential temperature values against height or pressure [10 marks]

   Is potential temperature constant with height? What does this say about the meteorological conditions?

   Compare your 'hill profile' with that obtained from a recent radiosonde ascent. Comment on how the profile you have obtained in ascending the hill will compare with a balloon ascent made in the `free' atmosphere.

2. Plot the values of q_v against height or pressure. [8 marks]

   Does this profile share any features in common with the potential temperature profile?

   Explain how the formation of a cloud over Goatfell (if observed) may have affected this profile.

Testing the hydrostatic relationship

3. [20 marks]
   Earlier we found that the hydrostatic equation may be written as

$$-Hp\ Ln\left(\frac{p}{p_0}\right) = z - z_0$$

where p is the pressure at any height z, p_0 is the pressure at a reference height z_0 and H_p is the scale height for pressure. (Note that the negative sign on the left hand side of this equation arises because the pressure decreases with height). Rearranging this equation gives

$$\ln p = \ln p_0 - \frac{z}{H_p} \; .$$

Plot a graph of ln(p) against z (*from the GPS observation*) fit a straight line through your data points (note that Microsoft Excel tends to round off the slope $1/H_p$ because it is a very small number. For this reason, it may be useful to plot ln(p) for the horizontal axis and z on the vertical axis, which gives a more precise value of H_p).

Recall the equation of a straight line: $y = mx+c$

Comparing with our hydrostatic equation, we see that if we plot ln(p) against z, then ln(p) is y, z is x, ln(p_0) is the intercept c on the y-axis and $-1/H_p$ is the gradient m (if you chose to plot z against ln(p) instead, then z is y, ln(p) is x, and $-H_p$ is m). From the gradient of the graph, determine H_p. For simplicity, first compare your value with the theoretical value in the absence of moist effects (i.e., with T_v in Eqn. (12) replaced by the temperature T). Note that T must be expressed in Kelvin, not °C, so add 273 to the temperature in °C.

Is your calculated value compatible with the theoretical one?

4. [8 marks]
Now compute the theoretical H_p with moist effects included by using the virtual temperature T_v.

Does the inclusion of moisture improve the agreement between the theoretical value of H_p *and the value that you determined graphically in exercise 3? (take note of error estimates when answering this question).*

5. [6 marks]
Now apply Equation (13) to estimate the altitude difference between the field centre and the summit of Goatfell. The pressure at Lochranza will change as you walk up and down the hill. Therefore we must estimate what the pressure was at Lochranza at the time when you made the observation at the top of hill.

Suppose that the pressures you measured at Lochranza were p_{start} when you started out at time t_{start} and p_{end} when you arrived back at Lochranza at time t_{end} (convert all times to minutes or seconds for this purpose).

Then suppose that at time t on Goatfell you measured pressure p_{meas}. Using linear interpolation, the corrected pressure p_{corr} is

$$p_{corr} = p_{meas} + \frac{(p_{end} - p_{start})}{(t_{end} - t_{start})}(t - t_{start})$$

You can now use Eq. (13) to calculate the altitude difference between Lochranza (z_1) and the summit of Goatfell (z_2) using the initial pressure at Lochranza (p_1) and the corrected pressure at the summit (p_2). Note that in applying Eq. (13) you will need to use the average absolute virtual temperature $\overline{T_v}$ between Lochranza and the summit.

What altitude do you get for the summit of Goatfell?

How accurate is your answer? What are your sources of error?

How could you use the existing data to improve this result?

Examining the influence of Goatfell on the winds

6. [10 marks]
From the Lochranza radiosonde launch on the morning of the hill walk, compute the "bulk" Froude number

$$Fr = \frac{U}{NH}$$

by taking H as the peak height of Goatfell, U as the mean observed flow speed between the surface and a height of about 2.5 km (or 750 mb), and

$$N^2 \approx \frac{2g}{\theta_1 + \theta_0} \frac{\theta_1 - \theta_0}{z_1 - z_0},$$

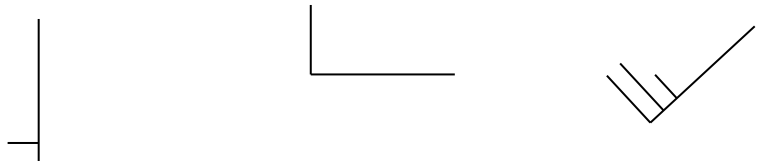
where θ_0 and z_0 are the potential temperature and surface elevation near sea level and θ_1 and z_1 are that at a height of around 2.5 km (or 750 mb).

Based on the direction of the low-level winds, comment on whether the air that you sampled during the hill walk was on the upwind or lee side of the mountain (or neither)?

Does this value suggest that the impinging flow was rising over the terrain or detouring around it during your ascent of Goatfell?

7. [10 marks]
Using the OS grid references and the wind measurements from the datasheet, draw wind barbs at each measurement location on the zoomed-in topographic map of the Goatfell route provided by the instructor. Use the examples below for guidance in drawing the wind barbs:

Southerly (180°), 2.5 m s^{-1} | Westerly (270°), 5 m s^{-1} | West-southwesterly (235°), 12.5 m s^{-1}



Compare the wind profile on the hill to that obtained from the Lochranza radiosonde launch on the morning of your hill walk.

Do the vertical variations in wind velocity measured during your ascent match those from the upstream sounding? Comment on the physical factors or processes that might account for the differences between them.

Wind and Energy Profile

Wind, Temperature and Energy profiles

Exchanges of Heat, Moisture and Momentum Between the Land Surface and the Atmosphere

Learning Outcomes

At the end of the exercises in this section, you will be able to:

- describe typical air temperature and wind speed profiles near the ground in different weather conditions and at different times of day
- understand how the various components of the surface energy balance can be measured
- calculate the albedo and emissivity of typical surface materials
- calculate fluxes of sensible and latent heat between the surface and the atmosphere
- relate surface energy balance to near-surface meteorology

Background Information

The atmospheric boundary layer

The lower part of the atmosphere, which is directly influenced by the surface of the Earth, is known as the *boundary layer*. It has a depth varying from tens of metres up to one or two kilometres in the UK. The boundary layer is important in meteorology for at least three main reasons:

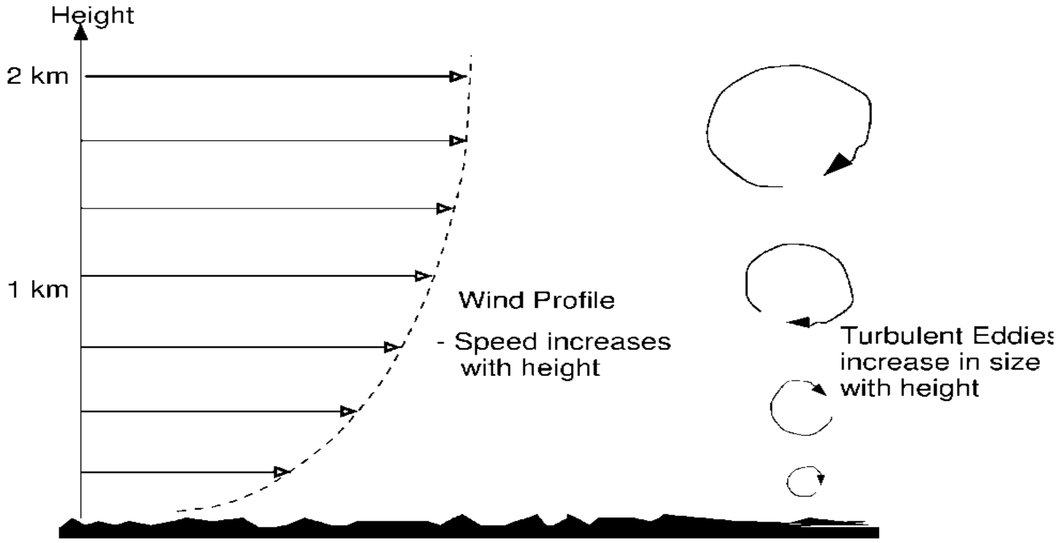
- It is the region in which most of the energy driving atmospheric processes originates as heating from the surface, which is warmed by radiation from the Sun.
- It is also the most important part of the atmosphere in the wider field of Environmental Science: it is where we live, produce our food and release the bulk of our anthropogenic pollution.
- The Earth surface (in particular the ocean surface) is the source for water vapour, which enters the air through evaporation into the boundary layer. The boundary layer is thus a crucial component in the hydrological cycle.

The nature of the boundary layer is somewhat different from that of the “free” atmosphere, in that boundary-layer dynamics are dominated by the occurrence of *turbulence*. Turbulence is random eddying motions of air which are set up by two primary physical processes:

- Radiative heating and cooling of the surface is stronger than at higher levels in the atmosphere, so the principal sources and sinks of heat are at the surface. Strong solar heating leads to turbulent convection and mixing in the boundary layer, whereas strong cooling at night tends to stabilise the boundary layer and suppress turbulence.
- Wind becomes turbulent as it flows over obstacles on the surface. This has important impacts on the transport of heat, moisture and chemicals from the surface to higher in the atmosphere.

Random motions of the air in turbulence are manifested by “gustiness” of the wind and are accompanied by fluctuations in temperature and humidity on similar timescales. The importance of turbulence is that it causes rapid diffusion, or mixing, of heat, moisture, trace gases and aerosols; turbulence effectively “stirs” the air close to the ground. The fact that turbulence also mixes wind momentum means that it transfers the frictional drag on air flowing over the ground into the upper air.

Turbulence is one of the great unsolved problems in modern science. Quantitatively, the way that we usually treat turbulence is to think of averaged quantities (such as the time-averaged horizontal wind speed) and then look at the statistics of variations around these values. For example, the standard deviation of wind speed about its mean value may be taken to be a measure of gustiness. Forecasting turbulence is important: gusts increase the potential of wind to damage trees and property, and turbulence is a hazard and occasionally fatal for aircraft.



Schematic diagram of the atmospheric boundary layer

Surface energy balance

In conditions of low wind, radiative transfer at the surface of the Earth dominates the production or suppression of turbulence. The heating of the lower metres of the atmosphere depends on the heating of the surface and the transfer of heat from the surface into the air. These processes can be quantified by looking at the *surface energy balance.*

The exchange of energy between the Earth surface and the overlying atmosphere involves four important processes:

- absorption and emission of natural electromagnetic radiation by the surface
- thermal conduction of heat energy within the ground
- turbulent transfer of heat energy towards or away from the surface within the atmosphere
- evaporation of water stored on or below the surface, or condensation of atmospheric water vapour onto the surface.

Each of these processes can be measured by an associated energy flux density, defined as the rate of transfer of energy across a surface of unit area. In SI, an energy flux density has units of J s^{-1} m^{-2}, which is the same as W m^{-2}. We will now consider each process in more detail.

The surface absorbs and reflects solar (or *shortwave*) radiation, and emits terrestrial (or *longwave*) radiation. *Solar radiation* refers to electromagnetic energy emitted by the Sun, which emits roughly as a *black body* with an emitting temperature of about 6000 K (energy emitted per unit surface area of a black body at Kelvin temperature T is σT^4, where $\sigma = 5.67 \times 10^{-8}$ W m^{-2} K^{-4}). Most of the energy is associated with the wavelength range 0.3 to 2.0 µm, with a maximum located near 0.5 µm in the *visible band* (0.4 to 0.7µm).

The *direct solar beam* can be considered as formed by parallel rays by the time it reaches the Earth's atmosphere. The *irradiance* of a surface normal to the beam outside the atmosphere is about 1380 Wm^{-2}, but *attenuation* reduces this to the order of 1000 W m^{-2} or less by the time the beam reaches the Earth surface. The surface also receives *diffuse solar radiation* scattered from air molecules, aerosol particles and clouds.

The *net solar irradiance* of the surface can be represented by

$$S_n = S_{\downarrow} - S_{\uparrow},$$

where $S_\downarrow$ and $S_\uparrow$ are the total down-welling and up-welling shortwave radiation, respectively. Because the Earth is not hot enough to emit shortwave radiation, $S_\uparrow$ is entirely associated with *reflection* of some of the down-welling radiation. It follows that

$$S_n = (1-\alpha)S_\downarrow,$$

where $\alpha = S_\uparrow/S_\downarrow$ is a reflection coefficient known as the *surface albedo*.

Terrestrial radiation refers to electromagnetic energy emitted by the Earth surface and the atmosphere. The Earth surface emits like a black body with an emitting temperature T_s of around 290 K. Most of the energy is associated with the wavelength range 3 to 30 µm and is often called longwave radiation. For most surface materials, the longwave energy emitted per unit area is predicted accurately by $\varepsilon_s \sigma T_s^4$, where ε_s is an *effective emissivity*. From *Kirchoff's Law*, it follows that a surface material *absorbs* a fraction ε_s of incident longwave radiation and reflects fraction $1-\varepsilon_s$. For down-welling longwave radiation $L_\downarrow$ from the atmosphere, the *net longwave irradiance* of the surface is therefore

$$L_n = L_\downarrow - L_\uparrow = L_\downarrow - [(1-\varepsilon_s)L_\downarrow + \varepsilon_s\sigma T_s^4] = \varepsilon_s(L_\downarrow - \sigma T_s^4).$$

If the surface is assumed to be a black body ($\varepsilon_s = 1$), this becomes the familiar $L_n = L_\downarrow - \sigma T_s^4$.

The atmosphere does not behave like a black body because its emissivity varies strongly with wavelength. It is almost transparent (ε very small) to radiation in the window region 8 to 12 µm but practically opaque (ε very large) in other parts of the spectrum (e.g. 3 to 8 µm and greater than 12 µm) due to absorption by *greenhouse gases* such as water vapour and carbon dioxide. Thick clouds act like black bodies with effective emissivities close to 1.

The *net surface irradiance R_n* is the sum of shortwave and longwave contributions:

$$R_n = (S_\downarrow + L_\downarrow) - (S_\uparrow + L_\uparrow) = S_n + L_n.$$

This represents the rate of *gain* of energy by the Earth through the absorption of radiation. It is defined to be a positive number when it is in the sense of warming the Earth's surface.

The radiative transfer of energy to the surface is only part of the story. The response in terms of heat flowing from the surface into the atmosphere has two components:

Sensible heat flux H: this is the upwards flux of energy into the air by conduction and convection ("sensible" because it can be directly sensed as a change in temperature). It depends on the temperature difference between the surface and the air, the wind speed and the strength of the turbulence (stronger turbulence increases the flow of heat). When warm air flows over a cold surface, H can be negative as sensible heat flows from the air into the ground.

Latent heat flux LE: this is the "latent" (hidden) flow of heat which would be released if water vapour evaporating from the surface at rate E condensed in the atmosphere. L is the latent heat of vaporization (2.45×10^6 J kg^{-1} at 20°C). LE can be negative when moist air flows over a cold surface and water vapour condenses as dew. Evaporation of water (or sublimation of ice) cools the ground and condensation warms it. Evaporation is generally large over open water or wet soil, and can include significant amounts of *evapotranspiration* from the leaves of plants.

There is also a significant conducted *ground heat flux G* due to temperature gradients below the surface. G is positive when it is warming and negative when it is cooling the ground.

Combining all of these terms, the energy balance for a surface layer of finite depth and unit horizontal area is

$$\frac{dQ}{dt} = R_n - G - H - LE,$$

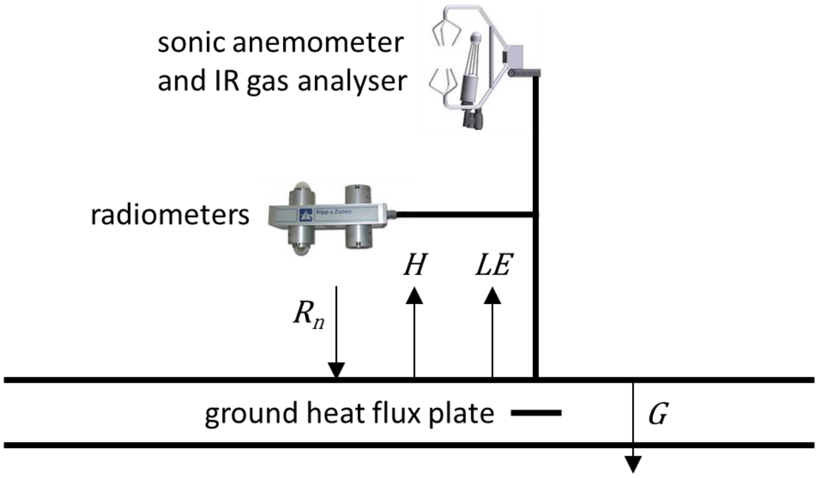
where Q is the total heat energy stored within the layer. In the special case of just considering an infinitesimally thin surface layer, there can be no heat storage and the *surface energy balance equation* reduces to

$$R_n - G - H - LE = 0$$

or

$$R_n - G = H + LE.$$

We will be making this assumption during our analyses. (Question: how realistic is this for soil or for an ocean?)



Schematic diagram of the measurement and balance of surface energy fluxes

The surface layer

Apart from the effect of surface heating on production and suppression of turbulence, there is also a significant process of turbulence generation by wind shear. The surface of the Earth is "rough" on all scales, from grains of sand and blades of grass to trees, buildings and mountains protruding into the air. As the wind flows over any of these obstacles, it produces turbulence on a similar horizontal scale to the size of the obstacle. Strong winds generate turbulence and produce a well-mixed boundary layer even at night under clear skies.

A standard result is that in neutral conditions (i.e. small vertical temperature gradients or strong winds) over homogeneous surfaces (i.e. no large horizontal variations in roughness), the average wind speed profile in the surface layer (up to about 30 m) has a logarithmic form

$$U(z) = \frac{u_*}{k} \ln\left(\frac{z}{z_0}\right),$$

where ln is the natural logarithm. The other parameters in this equation have the following meanings:

- z is height above the surface
- u_* is the *friction velocity* and is a measure of the strength of the turbulent variations in wind speed ("gusts") and the efficiency of the turbulence in mixing atmospheric constituents
- $k \approx 0.4$ is *von Kármán's constant*
- z_0 is the *roughness length* and is related to the size, shape and distribution of the surface obstacles over which the air is flowing.

Note that this logarithmic form fails to apply very close to the ground (where the function becomes infinite); in practice it is used down to around the roughness length for $z \gtrsim z_0$. There are departures from the logarithmic profile when the boundary layer is not neutral or the surface is not uniform.

The turbulent eddies in the boundary layer also carry heat away from the surface (when the ground is heated by the sun) or towards it (when the ground is cooled by radiation into space on a clear night). Under these circumstances, there is a temperature gradient in the surface layer that is closely related to the wind speed profile. A standard result is that when the heating or cooling is not too strong, the average temperature profile in the surface layer also has a logarithmic form

$$T(z) = T_s + \frac{T_*}{k} \ln\left(\frac{z}{z_0}\right),$$

where T_s is the temperature of the surface, T_* is the *friction temperature* and k has the same value as above.

If logarithmic profiles of temperature and wind speed are measured in the surface layer, the sensible heat flux can be determined from

$$H = -\rho c_p u_* T_*,$$

where ρ is the air density (typically about 1.25 kg m^{-3}) and $c_p = 1004$ J K^{-1} kg^{-1} is the specific heat capacity of air at constant pressure (check that this equation gives a flux in W m^{-2}).

Turbulent fluxes

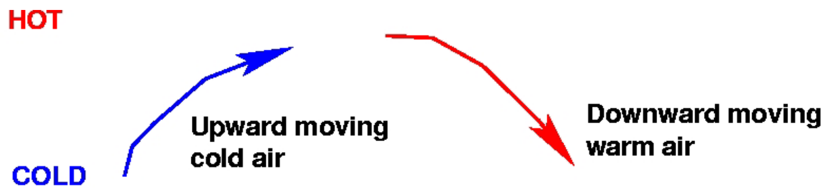
Sensible heat flux may also be interpreted in terms of the turbulent eddies. At a fixed point, turbulence may be thought of as apparently random fluctuations of all three wind components and temperature. If these fluctuations are measured fast enough, we can actually directly explain the effects of the eddies. An instrument to do this is the *sonic anemometer,* which you will be using in this exercise. Suppose that at an instant in time the temperature T at a point of interest is a little higher than average; we denote the temperature fluctuation by T', which is positive at this instant. In a turbulent flow, T' will sometimes be positive, sometimes negative and will average out to zero. Now suppose that the air is moving upwards at this instant (i.e. the vertical wind speed w is positive and, since the mean vertical wind is likely to be zero, w' is positive), so slightly warmer than average air is moving upwards. Thus we can say that at this instant, *heat is being transferred upwards*. However, if vertical motion and temperature are not related (we use the term *uncorrelated*) then at another instant when w' is positive, T' is just as likely to be negative (i.e. heat is being transferred downwards). Over many turbulent eddies, the heat transfer will average out to zero.

In contrast to the above, suppose that the air is warmer close to the ground than at higher levels. We then expect that upward moving air will be systematically warmer (i.e. positive T' along with positive w') than downward moving air (which will have negative T' and negative w'). Averaged over many turbulent eddies, there will be a net upward transfer of heat; we call this a *turbulent heat flux*. Mathematically, if T' and w' are either both positive or both negative, then $w'T'$ is positive. We can determine the heat flux by taking an average of $w'T'$ over many turbulent eddies.

Using this argument, the sensible heat flux is given by

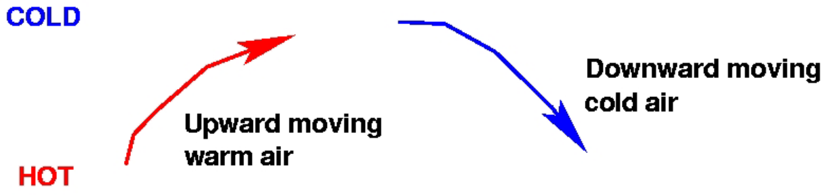
$$H = \rho c_p \overline{w'T'},$$

where the bar over $w'T'$ means that a quantity is averaged in time. This is known as the *eddy covariance method* because averaging the product of fluctuations in w and T gives their covariance. Graphical illustrations of the heat transfer process are shown on the next page.



Downward Heat Flux

If upward moving air ($w' > 0$) is generally colder than average ($T' < 0$), w' and T' have opposite signs. An average of $w'T'$ over many measurements will be negative – a downward heat flux.



Upward Heat Flux

If upward moving air ($w' > 0$) is generally warmer than average ($T' > 0$), w' and T' have the same sign. An average of $w'T'$ over many measurements will be positive – an upward heat flux.

The same arguments can be applied to upward or downward transfers of momentum and water vapour. If u' and v' are fluctuations in the horizontal wind components, then an upward or downward transfer of faster moving air (a *turbulent momentum flux*) corresponds to systematic relationships between u' and w' and/or v' and w'. The quantities $\rho\overline{u'w'}$ and $\rho\overline{v'w'}$ are the two vector components of the vertical flux of horizontal momentum and are related to the friction velocity by

$$u_* = \left[\left(\overline{u'w'}\right)^2 + \left(\overline{v'w'}\right)^2\right]^{1/4}.$$

Similarly, if the water vapour content of the air measured by specific humidity q increases or decreases with height, then turbulence will carry a moisture flux $E = \rho\overline{w'q'}$.

The Penman-Monteith equation

Rather than having to obtain accurate and rapid-response measurements of humidity for the eddy covariance method or accurate measurements of humidity profiles for the profile method, as described above for sensible heat fluxes, latent heat fluxes can be approximated using the *Penman-Monteith equation*

$$LE = \frac{\frac{\rho c_p}{r_a}[e_s(T) - e_a] + \Delta(R_n - G)}{\Delta + \gamma(1 + r_s/r_a)},$$

where e_a is the pressure of water vapour in the air, $e_s(T)$ is the vapour pressure for saturation at temperature T, $\Delta = de_s/dT$, $\gamma \approx 67$ Pa K^{-1} is the *psychrometer constant* (actually a function of air pressure rather than a constant), and r_s and r_a are resistances for moisture transport through the surface and the air, respectively. This only requires measurements of average temperature, humidity and wind speed at a single height above the surface, plus net radiation and ground heat flux.

You do not need to know a full derivation of the Penman-Monteith equation, which relies on the surface energy balance equation, logarithmic profiles and assumptions about the wetness of the surface. Instead, here is an interpretation of the components of the equation:

- the first term in the numerator relates to the rate of evaporation that occurs if the surface is saturated and has the same temperature as the air
- the second term in the numerator is a correction for surface temperature taking account of net radiation and ground heat flux
- the denominator corrects for the fact that some of the additional energy flow generated by correcting the surface temperature is realised as sensible heat flux.

Overview of exercises

The instrumentation in the field includes:

- a 10 m mast with measurements of temperature and wind speed at 6 heights, a measurement of humidity at one height, a buried heat flux plate, a sonic anemometer and an infrared gas analyser (IRGA) for measuring water vapour and CO_2
- an automatic weather station recording standard meteorological variables
- a radiometer mast measuring the four components $L_{\downarrow}$, $L_{\uparrow}$, $S_{\downarrow}$ and $S_{\uparrow}$ of surface radiation from which R_n can be calculated
- a portable four-component radiometer system.

You will use the portable radiometer system in the field to measure the albedo and emissivity of different surfaces in the ALB exercise. The SEE and SLD exercises will begin with a tour of the field to discuss the measurement principles used by the instruments, followed by classroom activities using the data.

As described above, sensible heat flux can be calculated by three different methods:

- the temperature and wind speed profiles give $H = -\rho c_p u_* T_*$
- the sonic anemometer gives $H = \rho c_p \overline{w'T'}$
- the Penman-Monteith equation and the surface energy balance give $H = R_n - G - LE$.

The SLD exercise uses knowledge of turbulent surface layer dynamics to calculate H by the profile and eddy covariance methods. Combining data from the sonic anemometer and the IRGA additionally allows calculations of evaporation and CO_2 fluxes from the surface. The SEE exercise uses surface energy balance and the Penman-Monteith equation to calculate H and compares results from the three different methods. Using data from the weather station you will be able to relate the evolution of the components of the surface energy balance to the meteorology we have experienced during the course.

Because the SEE and SLD exercises use data from a variety of different instruments, you should make sure that measurements are available from all of the instruments for the time periods you wish to study. There are many subtasks in these exercises and you should consider how to divide the work up among your group.

ALB Exercise: Albedo and Emissivity of Surfaces

MARKS

- 50% for performance in field
- 50% for analysis

GROUP AND INDIVIDUAL WORK REQUIREMENTS

- You may share raw data from the instruments and field notes with the whole class.
- You may share tables/plots of data with your specific group.
- You must carry out the analysis of the data and answer the questions yourself as an individual. You must write up the answers to the questions in your own words.

Goals

- measure upward and downward shortwave and longwave radiation components, ground temperature and effective black body temperature over several different surfaces (e.g. grass, gravel, concrete, shrubs, bare soil)
- estimate the albedo of each surface
- estimate the emissivity of each surface
- assess the uncertainty and error in your experiment
- use your results to estimate the planetary albedo

Preparation

This is a team effort. You will design the experiment in the class room together with all of the groups before carrying out your part of the activity. Make sure you read the preceding information on surface energy balance and errors, and that you carefully understand the experiment design before embarking on the fieldwork. You must come to the field with a detailed plan for what you need to do in your group and how you will do it, although as with any field campaign, you must also be ready to adapt quickly to changing conditions and information. Read the sections below so that you can make a good plan with the whole class.

Equipment

The kit for this experiment consists of four radiometers mounted on a camera tripod and attached to a Campbell data logger. You will also have an infrared thermometer and a handheld thermometer for measuring surface temperature. The radiometers are delicate and expensive pieces of equipment, so please be careful with them. Assume that the uncertainty in their measured radiation fluxes is 5%. The tripod and radiometers are heavy, so please be careful when lifting them. The two radiometers with glass domes measure shortwave radiation, whereas the two with flat plates measure longwave radiation (Q: why are there only domes on the shortwave radiometers?). It is important when you make your measurements that the tripod is on level ground and the radiometers are aligned in a horizontal plane. If it has been or is raining, it is also important that there is no water on the surfaces of the radiometers (Q: what effect would this have on the radiation you measure?). If the surfaces/domes are wet, you can ask the activity leader to dry them gently with a tissue. It will be best to avoid doing this practical in heavy rain if at all possible.

Output voltages produced by each radiometer are converted to radiation fluxes in Wm^{-2}. Values are measured every second and averaged over a minute by the data logger. Data will be recorded by the logger continuously throughout the experiment, so you will need to note the start and end time of each of your sample periods. At the end of the experiment, data can be downloaded via the 232 port on the logger to the laptop provided; please ask the activity leader to do this for you. Your data can then be transferred via a memory stick to the PCs at your table in the classroom for analysis.

Methodology

1. Position the radiometers over a surface of interest (start with a fairly uniform patch of grass). Make sure that the arm is level and flat (use the spirit level on the radiometer arm) with one dome pointing up and one down. Check that the radiometers are dry.
2. Begin your first IOP (intensive observing period). Collect data from the radiometers for at least 10 minutes (the IOP length needs to be decided in advance by the whole class as part of the experiment design). During this period you should measure the surface temperature (T_s) several times using the handheld probe and the effective black body temperature of the surface (T_{IR}) using the infrared thermometer. Also note any change in meteorological conditions during this time.
3. After the IOP has finished, ask the activity leader to cover the radiometers with the black bag, carefully move the tripod (and logger if necessary) to a position over a different surface or move the surface under the radiometers, and start a new IOP by removing the bag and noting the time.
4. Repeat the IOP for the different materials you have been assigned by the class, which you should choose to represent different surfaces found on Earth. You can fill a metal tray with material for some of these case studies. If you use a tray, make sure that the height of the radiometer arm is 50 cm above the surface and that your tray lies on a surface which you have already measured the properties over, as this will affect your results (Q: why?). If you are using a tray, consider the influence of the tray material and what tests you might do to examine this effect.
5. Inform the activity leader that you have completed your IOPs and request that they collect the data from the logger. With their help, upload the data from the logger to the laptop and then to a memory stick. This memory stick can be taken back to the classroom to load the data onto your PCs. **Please ensure that the memory stick is returned to the activity leader.**
6. Collate your measurements into data tables that can be used by the whole class, use one table for each IOP (you will need to have agreed a format for the tables with the other groups to make sure that all the required information is present). When the table is complete, upload it to the shared drive so that all other groups can access the data. Do this as soon as possible after carrying out the experiment.
7. When all groups have uploaded their data, you can complete the analysis (below).

Analysis

Tip: In the following analysis, make sure you report your raw data from the shared data tables, as well as any processed data and your calculated values. You should explain any approximations or assumptions that you make, and show your full workings for one example of your calculations for α, α_{meas}, and ε_s.

1. Using measurements of incoming shortwave radiation $S_\downarrow$ and reflected shortwave radiation $S_\uparrow$, calculate the albedo $\alpha = S_\uparrow / S_\downarrow$ for all of the surfaces. For experiments in which you used a tray, you will also need to consider that the field of view of the radiometers is larger than just the tray itself. This can be done using the equation:

$$\alpha_{meas} = w_g \alpha_g + (1 - w_g)\alpha_t$$

   where α_{meas} is the measured albedo, w_g is the fraction of irradiance coming from outside the tray (0.78 ± 0.02 when the tripod arm is 50 cm high), α_g is the albedo of the ground that the tray is placed on and α_t is the albedo of the material in the tray.

   [5 marks]

2. Calculate the emissivity ε_s for each surface from the equation

$$L_\uparrow = \varepsilon_s \sigma T_s^4 + (1 - \varepsilon_s) L_\downarrow$$

   using measurements of incoming longwave radiation $L_\downarrow$, outgoing longwave radiation $L_\uparrow$ and surface temperature T_s in each case ($\sigma = 5.67 \times 10^{-8}$ Wm^{-2} K^{-4}).

   [5 marks]

3. Comment on the quality of your measurements in Q1 and Q2. You may wish to:
   a. Explain whether or not you excluded any data from your analysis
   b. List the possible sources and types of errors.
   c. Rank the sources of error in order of their contribution to the total error (which of these errors are the most important to account for and which are negligible?).
   d. Explain any steps taken to minimise error.

   [20 marks]

4. Compare your T_s and T_{IR} measurements for different surfaces. The T_{IR} measurement assumes that the surface is a blackbody ($\varepsilon_s = 1$), and therefore the equation $L_\uparrow = \sigma T_{IR}^4$ should hold. Is this true in your data? If not, why might this be?

   [5 marks]

5. Briefly describe what your results show about the different materials you have tested. Do you think that your surfaces are a good representation of different surfaces around the Earth? If not, why not?

   [5 marks]

6. Use your results to calculate the planetary albedo of the Earth with and without clouds. You might like to consider the fraction of the Earth covered by land, ocean and other surfaces, and assume that around 70% of the Earth is cloud-covered on average. Explain your reasoning and calculations.

   [10 marks]

SEE Exercise: Surface Energy Exchange

All marks for this exercise are from answers in your notebook

Group and individual work requirements

- you can share plots and tables of data within your group
- you can discuss interpretations of the data within your group
- you must write up answers to the questions in your own words

Goals

- calculate sensible and latent heat fluxes from meteorological variables and surface energy balance
- describe variations in components of the surface energy balance over time
- relate surface energy balance to changes in meteorological conditions
- compare results of flux calculations by different methods

Methodology

The activity leader will show you where to find the file of data collected every minute to date by the meteorological mast. Make a copy of this file and open it to see that it contains a row of data for each minute, preceded by several header rows which describe the data in each column. Take some time to familiarize yourself with the data (read the headers and try plotting some of the data in columns to see if they make sense).

A spreadsheet `SEE.xlsx` which calculates the components of the surface energy balance using the Penman-Monteith equation will be provided for you to copy. The spreadsheet contains headers but no data when you first open it. Not all of the data measured by the mast are required for this exercise; copy the required variables from the mast file to the corresponding columns in the spreadsheet. Once data has been entered in `SEE.xlsx`, numbers will appear in row 5 of columns O to W (scroll right to see them all). Select these cells and drag to copy the equations that they contain over the same number of rows as the entered meteorological data.

The spreadsheet contains equations to calculate humidity variables, plus a table of physical constants and a table of adjustable parameters. These are used along with temperature, relative humidity and wind speed measured at one height on the mast to calculate latent heat flux LE according to the Penman-Monteith equation. Then the sensible heat flux H is calculated using measured net radiation R_n and ground heat flux G in the surface energy balance equation.

1. Plot a graph showing time series of G, H, LE and R_n all on the same axes. Discuss the graph within your group and with the activity leader to choose a day with interesting features to examine more closely. Describe variations in surface energy fluxes with relation to the meteorological conditions.

   [10 marks]

2. What do the relative sizes of H and LE during this day imply about the surface moisture? Increase and decrease the surface resistance r_s from the initial value of 60 s m^{-1} in cell L11 to see what influence it has (surface roughness length z_0 in cell L12 can also be adjusted).

   [10 marks]

3. Calculate the average latent heat flux over the day. Taking the latent heat of vaporization of water to be $L = 2.45\times10^6$ J kg^{-1}, what is the average rate of evaporation in mm per day?

   [10 marks]

The next two questions can be answered once you have results from the SLD exercise.

4. For the two 1-hour periods you have used in the SLD exercise, record averages of u_*, H and LE from `SEE.xlsx` in a table. Add 1-hour averages from the profile and eddy

covariance methods to the table (note that there is no calculation of LE from the profile method because a humidity profile has not been measured).

[10 marks]

5. Discuss how well the estimates from each method compare and why they might differ. What assumptions are made by each method and why might they not be valid?

[10 marks]

The spreadsheet does all of the surface energy balance calculations for you, but the equations that it uses are given below for information.

Saturated vapour pressure e_s (hPa) for measured air temperature T (°C) at height z_T (m) is estimated by the Tetens' formula

$$e_s = 6.11 \exp\left(\frac{17.27\,T}{T + 237.3}\right).$$

Using this, the measured relative humidity RH (%) is converted to vapour pressure e_a (hPa) using

$$e_a = \frac{RH}{100} e_s\,,$$

and the slope of the saturation vapour pressure curve is given by the Clausius-Clapeyron equation

$$\Delta \equiv \frac{de_s}{dT} = \frac{Le_s}{R_v(T + T_0)^2}\,,$$

where $L = 2.45 \times 10^6$ J kg^{-1} is the latent heat of vaporization, $R_v = 461$ J K^{-1} kg^{-1} is the gas constant for water vapour and $T_0 = 273.15$ K is the conversion from Celsius to Kelvin. Assuming logarithmic profiles, the friction velocity for measured wind speed U (m s^{-1}) at height z_U (m) is given by

$$u_* = \frac{kU}{\ln(z_U/z_0)}$$

and the aerodynamic resistance by

$$r_a = \frac{\ln(z_T/z_0)}{ku_*}$$

for roughness length z_0 (m). Calculated values are plugged into the Penman-Monteith equation with measured net radiation R_n and ground heat flux G (both in W m^{-2}) to give latent heat flux

$$LE = \frac{\frac{\rho c_p}{r_a}[e_s(T) - e_a] + \Delta(R_n - G)}{\Delta + \gamma(1 + r_s/r_a)}\,,$$

where $c_p = 1004$ J K^{-1} kg^{-1} is the specific heat capacity of air at constant pressure, air density ρ is taken to be 1.25 kg m^{-3} (it can be calculated more precisely from the ideal gas law if air pressure is also known) and $\gamma = 0.67$ hPa K^{-1} is the psychrometer constant. The surface resistance r_s is 0 s m^{-1} if the surface is saturated and large if the surface and the soil are dry. Finally, the calculated latent heat flux is used with the surface energy balance equation to find sensible heat flux

$$H = R_n - G - LE.$$

SLD Exercise: Turbulent Fluxes in the Surface Layer

All marks for this exercise are from answers in your notebook

Group and individual work requirements

- you can share plots and tables of data within your group
- you can discuss interpretations of the data within your group
- you must write up answers to the questions in your own words

Goals

- measure temperature and wind speed profiles near the ground
- relate measurements to theoretical profiles
- calculate friction velocity and sensible heat flux using the profile method
- calculate sensible heat and latent heat fluxes from eddy covariance measurements

Profile methodology

Temperature sensors and cup anemometers are mounted at six heights on the meteorological mast to measure temperature and wind speed profiles near the surface. The heights of the mounting arms are given in this table:

level	1	2	3	4	5	6
height (m)	0.5	1.0	2.0	3.5	6.3	9.1

Thermometers are offset by –0.08 m relative to these heights and anemometers by +0.12 m.

The profile method for calculating turbulent heat fluxes uses differences between temperatures measured at different heights in the atmosphere. Because the temperature difference between the top and bottom of the 10 m mast is expected to be small, systematic errors in the data from any of the temperature sensors will make the profiles useless. This problem can be compensated for by applying a temperature offset to each of the sensors based on cross-calibration. All six temperature sensors will have been run alongside each other for several hours before the mast was erected to reveal any biases. You will be provided with corrected data, but you can also access the uncorrected data from the calibration period if you would like to use it to estimate residual errors in the temperature measurements.

1) Select two hours of data for detailed analyses: an hour when the temperature profile was stable (temperature increases with height) and an hour when the profile was unstable (temperature decreases with height). When are these conditions expected? You may find it easier to identify stable and unstable hours by plotting time series of temperature *differences* between levels on the mast, rather than the temperatures themselves. You should also check that sonic anemometer and radiometer data are available for these hours for comparison with results from the SEE exercise. Explain your choices.

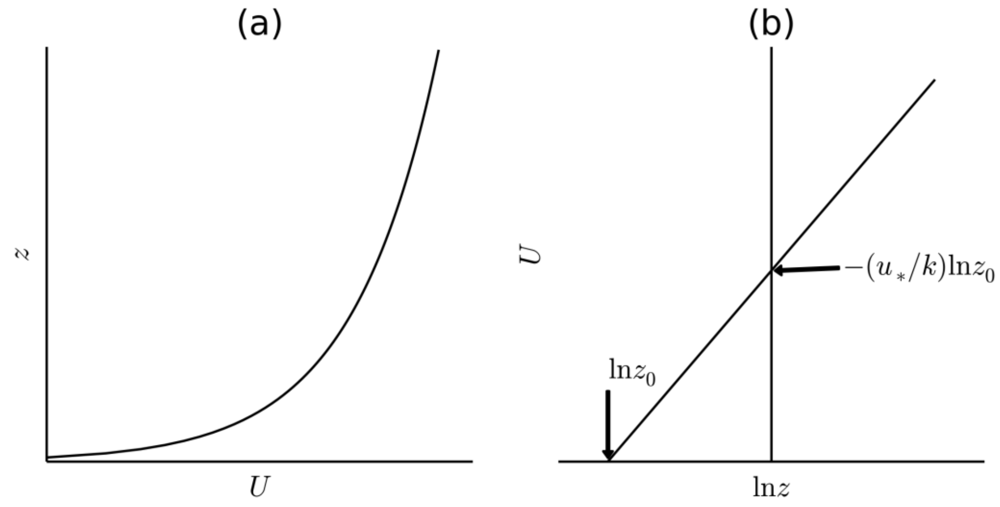
   [10 marks]

2) The anemometers output a pulse for every revolution of the cups, and the data logger converts the number of revolutions per minute into a wind speed (in m s^{-1}) using the manufacturer's calibration. Calculate means and standard deviations of wind speed separately at each height and for both selected hours. Plot graphs of mean wind speed against height, including error bars.

   [10 marks]

3) Plot wind speed in m s^{-1} on the y axis against ln z (natural logarithm of height) on the x axis of a graph for each selected hour. Fit straight lines through the data points.

   [10 marks]



Logarithmic wind profiles plotted with
(a) U on the x axis and z on the y axis, (b) ln z on the x axis and U on the y axis

From the theoretical logarithmic profile, wind speed U is related to height z above the ground:

$$U(z) = \frac{u_*}{k} \ln\left(\frac{z}{z_0}\right).$$

This is equivalent to

$$U(z) = \frac{u_*}{k} \ln z - \frac{u_*}{k} \ln z_0.$$

If we plot a graph of U against $\ln z$ and fit a straight line of the form $U = m \ln x + c$, then we expect that the gradient of the line is

$$m = \frac{u_*}{k}$$

and the intercept on the y axis is

$$c = -\frac{u_*}{k} \ln z_0.$$

4) For each selected hour, calculate u_* from the gradient of the fitted line and z_0 from the intercept on the y axis. Theoretically, a plot of U against $\ln z$ is expected to be a perfect straight line. In reality, it is clearly not a perfect fit. Discuss why this may be so, considering both the measurements themselves and the relevance of the logarithmic profile for the particular measurement site and times.

[10 marks]

Temperature profiles can be analysed in the same way. The theoretical profile

$$T(z) = T_s + \frac{T_*}{k} \ln\left(\frac{z}{z_0}\right)$$

can be rearranged as

$$T(z) = \frac{T_*}{k} \ln z + \left(T_s - \frac{T_*}{k} \ln z_0\right),$$

so if we plot a graph of T against $\ln z$, we expect to get a straight line with gradient

$$m = \frac{T_*}{k}$$

and intercept on the y axis

$$c = T_s - \frac{T_*}{k} \ln z_0.$$

5) Plot T against $\ln z$ and fit a straight line for each hour. Calculate T_* from the gradient of the straight line. Calculate the surface temperature T_s from the intercept on the y axis using T_* and values of z_0 from question 4 (surface temperature can also be calculated from longwave radiation as in the ALB exercise). Calculate the surface sensible heat flux H for each of the selected hours using

$$H = -\rho c_p u_* T_*$$

where ρ is the air density (typically about 1.25 kg m^{-3}) and c_p is the specific heat capacity of air at constant pressure (1004 J K^{-1} kg^{-1}).

[10 marks]

The SEE exercise includes a synthesis of results obtained by the profile method, eddy covariance and the Penman-Monteith equation.

Eddy covariance methodology

The mechanical inertia of a cup anemometer means that it may fail to turn at low wind speeds and may continue spinning after the wind has dropped from high speeds. The thermal inertia of a thermometer means that it takes some time to respond to changes in temperature. In place of the wet-bulb thermometer traditionally used for measuring humidity, automatic weather stations commonly use *humicaps* (humidity capacitors); these, again, are slow to respond to fluctuations. All of these instruments are fine for measuring average conditions over a few minutes, but instruments with much faster response times are required to measure turbulent fluctuations and calculate fluxes using the eddy covariance method.

A sonic anemometer has no moving parts. Short pulses of ultrasonic sound are exchanged along three different axes by pairs of transducers in a non-orthogonal arrangement to minimize flow distortion. The transit time depends on the speed of sound and the component of the wind vector along the axis, both of which can be found by setting each transducer alternately to transmit and receive. The speed of sound in air depends on temperature and slightly on humidity, so a sonic anemometer can also be used to make fast measurements of virtual temperature.

An infrared gas analyser measures the concentration of trace gases in the air by determining the absorption of radiation from an infrared source. In the LI-COR gas analyser that we are using, radiation from a broadband source is filtered to select wavelengths of 2.7 μm absorbed by water vapour and 4.3 μm absorbed by CO_2, so concentrations of both gases can be measured by the same instrument.

The Metek sonic anemometer can measure the three components of the wind vector (u, v, w) and air temperature (T) and the LI-COR infrared gas analyser can measure concentrations of water vapour (q in kg of water vapour per kg of air) and CO_2 (c in μmol per mol of air) many times a second. For this experiment, they are set to run at a relatively low rate of 5 Hz. Even so, a huge amount of data will be produced over the course of the week; you can access the raw data in the `rotated` folder if interested, but you will also be provided with processed hourly data. Extract H, LE and u_* for the hours you used in the profile calculations and copy them to a table for SEE question 4.

Balloon Profiling

Wind profiling with balloons

MARKS
- 20% for performance in field
- 80% for analysis

GROUP AND INDIVIDUAL WORK REQUIREMENTS
- You may share raw data from the instruments and field notes with the whole class.
- You may share tables/plots of data with the whole class.
- You must carry out the analysis of the data and answer the questions yourself as an individual. You must write up the answers to the questions in your own words.

Learning Outcomes

At the end of balloon profiling, you will be able to:
- explain features in the local terrain-induced air flow by analysing balloon trajectories,
- understand the valley air flows at different times of day and under different conditions,
- compare data at different meteorological scales,
- analyse complex, noisy data, to check for consistency,
- communicate effectively with other teams in the field.

Background Information

In the troposphere, windspeed and direction vary strongly as a function of height – from zero windspeed due to the friction of the surface, to fast jet stream flows which govern the pattern of weather system in the mid-latitudes. The temperature structure of the boundary layer overnight can lead to nocturnal jets, and lateral heating of the ground at synoptic scales can lead to the thermal wind. Good observations of the wind profile are needed for many applications: weather forecasting, aviation, pollution modelling, wind energy predictions, designing buildings, etc.

A simple model of the wind profile within the surface layer of the boundary layer is given by the logarithmic profile (see section on "wind and energy profiling"). This model is valid for nice flat terrain with uniform surface type, i.e. for homogeneous terrain. Arran is not very homogeneous! Here are some examples of how heterogeneity in surface type, elevation and heating can affect the wind profile:

Sea (and land) breezes: the temperature of any water mass varies far less than adjacent land under the same diurnal variation of solar radiation (think: why is this?). As a result, the buoyant air parcels over land will rise rapidly, inducing a flow from water to land at the surface, and a

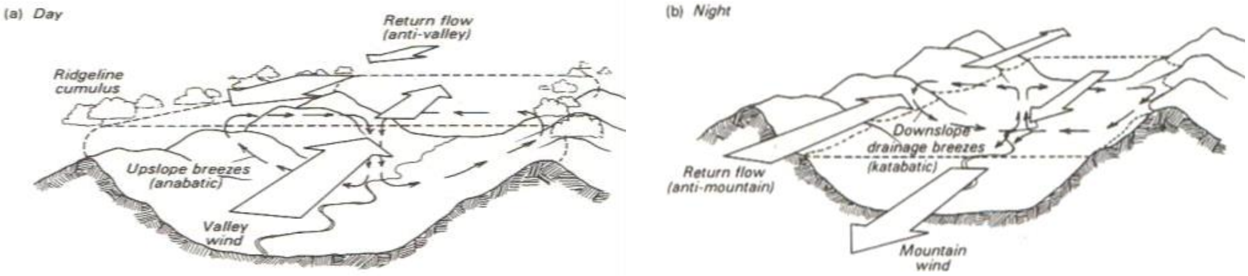


Figure 2: Valley winds by a) day and b) night

return flow aloft. Sea breezes contain very dense, cold air which can flow inland for many kilometres once initiated, due to negative buoyancy – a so-called “density current”. Convergence of sea breezes due to an irregular coastline can lead to uplift and localised convection, and a line of cumulus clouds can form. At night, the land cools more quickly than the water, and the flow reverses – land breezes tend to be weaker as the temperature contrast is not so stark.

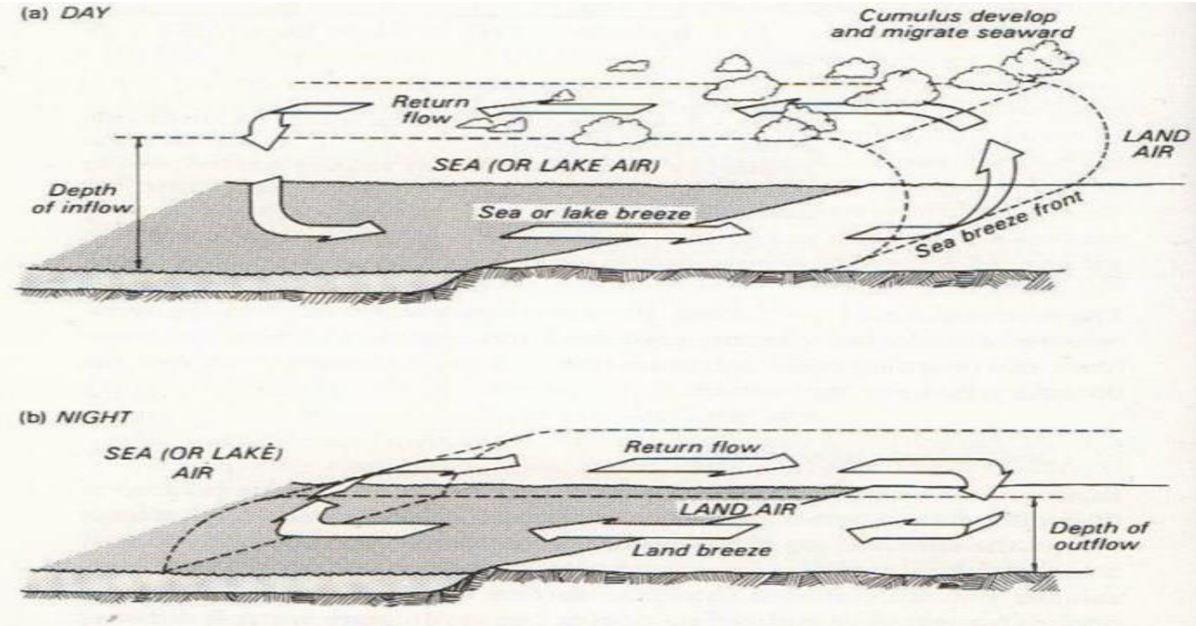
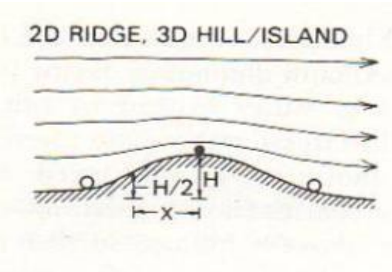


Figure 1: Sea breeze structure by a) day, and b) land breeze by night.

<u>Valley flow (anabatic and katabatic winds):</u> another buoyancy driven flow is due to intense heating of valley slopes which are well exposed to the sun. As air above the sunlit slopes starts to rise, flow is induced up the valley slope (*think: what does this have to do with mass continuity?*). Upslope flows are called “anabatic”. Once night falls, the tops of the hills cool rapidly as there is a large sky view factor. Cold, dense air starts to flow down the valley slopes, causing “katabatic” flows and cold pools of air can stagnate in local dips in the topography. Such “frost hollows” become very stably stratified with strong temperature inversions.

<u>Flow over shallow and steep hills:</u> an important phenomenon for flow over hills is *flow separation*. Normally, air flows over a surface experiencing friction with it due to a combination of molecular viscosity and turbulence. When the slope of a hill is small, flow “adheres” to the surface. The streamlines (along which a balloon should travel!) constrict, indicating that the flow “speeds up” (think: where on a hill to you tend to see wind turbines?). If the slope becomes too steep, the flow undergoes separation: flow “overshoots” at the top as inertial forces overcome viscous forces, a pressure gradient ensues, vorticity is generated and thus a “lee eddy” is formed. This area of the flow is highly turbulent (very dangerous for light aircraft!). The downstream extent of the eddy depends on the slope and height of the hill, and how turbulent conditions are: more turbulent conditions lead to a shorter wake region. (think: stability plays a role too... see “hill profile” section).

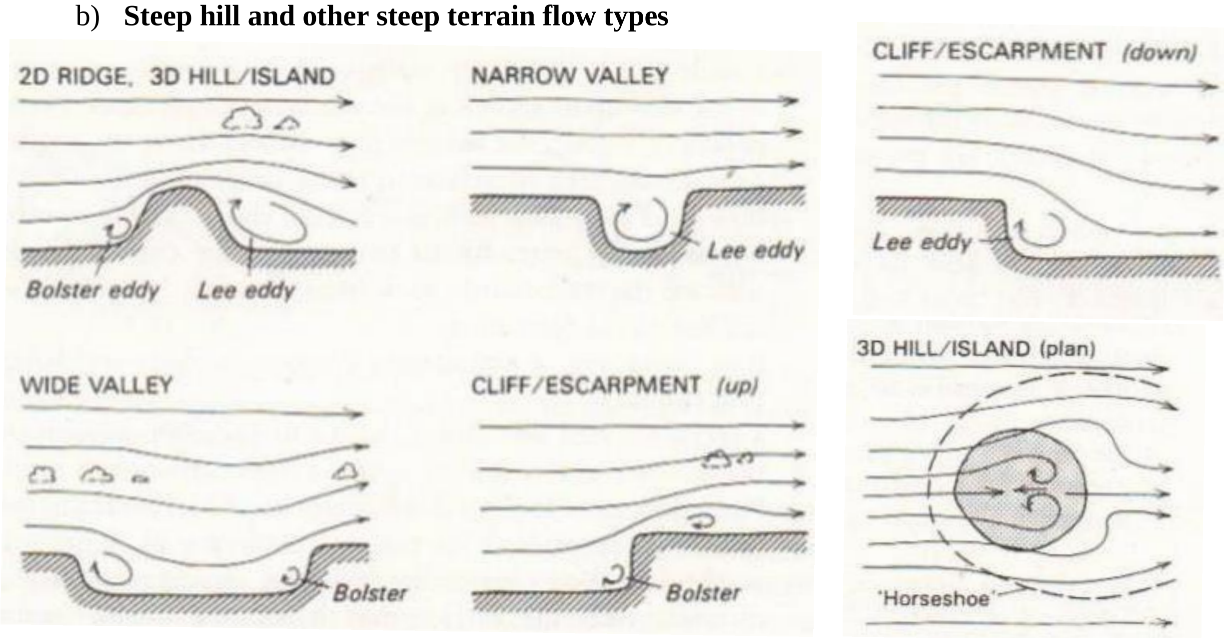
Figure 3: Flow over shallow and steep terrain. u_{up} is the upstream windspeed, u_{max} is the speed-up.



a) Shallow hill

Speed-up equation: $\frac{\overline{u_{max}}}{\overline{u_{up}}} \sim 1 + 1.8H/X$

b) **Steep hill and other steep terrain flow types**



<u>Boundary layer structure:</u> in a boundary layer over homogeneous terrain, the wind responds to the thermal structure, and also is a result of a balance between friction, Coriolis and pressure gradient forces. When there is no strong heating or cooling (neutral conditions), an "Ekman spiral" forms, where the wind veers with height (in the northern hemisphere):

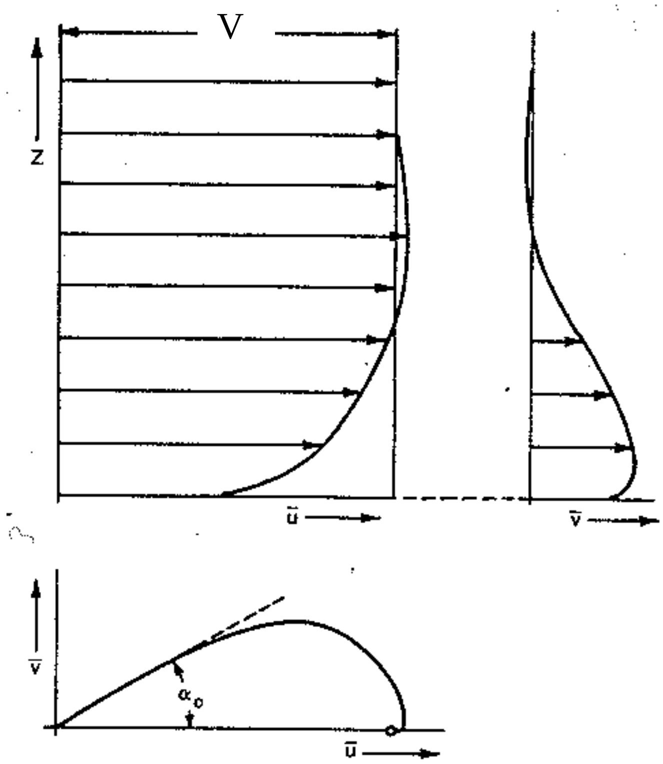
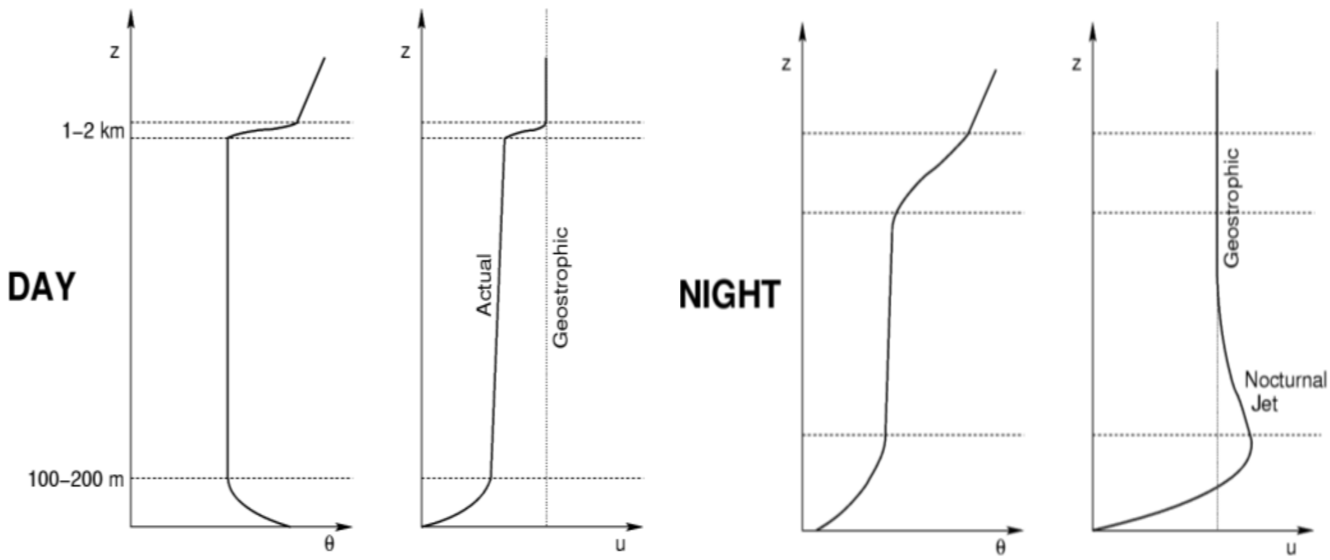


Figure 4: (above) Profiles of horizontal wind components u and v in Ekman boundary layer (below) hodograph showing angle α between boundary layer wind and geostrophic wind direction (along x-axis).

(*think: what is the balance of forces at the bottom and top of the profile? Why might this cause the wind direction to change?)*

When the surface heats by day, air is mixed vigorously by thermals rising from the surface up to the boundary layer inversion. Note the structure of the winds: little variation with height in the bulk of the boundary layer, and some shear across the inversion. When the surface cools, turbulence dies near the surface and the air no longer exchanges momentum with the surface – the flow "feels" no friction with it. This can lead to acceleration aloft, and a "nocturnal jet" can form just above the surface inversion, i.e. normally between 100 and 500m height (think: is this good or bad news for a wind turbine?).



Balloon tracking experiment

You will determine windspeed and direction as a function of height and location, to test whether the profile can be described simply in terms of surface friction, or is responding to local variations in topography, surface cover or surface heating.

The equipment consists of two theodolites, used for tracking the flight of pilot balloons. The balloons are filled with gas to make them slightly buoyant. The theodolites are used to record the *azimuth* and *elevation* angles of each balloon's flight. By making simultaneous measurements with two theodolites, the horizontal location and height of the balloon can be determined by triangulation.

WARNING: This practical requires intense teamwork!! Good luck...

Method

Form two groups of four people, each group working with a theodolite. Great care is needed in the initial set-up of the theodolites, and when taking measurements, in order to acquire useful data. The two groups ***must*** make their measurements synchronously and at carefully timed intervals: stopwatches are available to help with timing and two-way radios (or hand signals!) can help communication. Make a careful note of all the measurements detailed below. Double check that everything that needs to be recorded has been after each release. The theodolites will have already been positioned in the field for you.

As the balloons are released, keep a note of the time of the release, and track them for as long as possible, taking measurements at predefined times with as short an interval between measurements as you can. You should aim for a measurement about every 15 seconds, but may need to start with a longer interval and decrease as you gain practice. At each sample interval, record the time (in seconds), and the azimuth and elevation angles. Be sure to make a note of the sample at which magnification is switched from low to high power.

Within the group, ensure that each person plays a different role: Theodolite Reader (x2), Timer, Scribe and observer (to note changes in weather/wind conditions and any unusual events during ascent).

Your record of each release should look something like this....

Pilot Balloon Tracking Form

Release Point: Launch No:

Weights: Balloon type:

Date: Time of launch GMT:

Azimuth of north:

Azimuth of other theodolite:

Position of theodolite

Weather conditions:

Time	Azimuth	Elevation	Magnification : Hi/Low?

Data analysis

Once you are back in the classroom, please transfer your measurements to the pre-prepared excel spreadsheet (ask a member of staff for the location of this). Once each of the two groups involved has transferred their measurements let a member of staff know and they will process the data (or show you how to do so) to produce plots as shown below for each ascent. While this is taking place, you should review the material below which describes how the measurements are used to determine the balloon position and altitude and, ultimately, its velocity.

Geometrical method for determining balloon position

Having made observations of the azimuth and elevation of a pilot balloon from each of two theodolites, simple trigonometry allows the altitude and horizontal position of the pilot balloon to be calculated. This technique was used for many years to obtain wind profiles for aviation and weather forecasting, along with related techniques using a single theodolite. Today wind profiles are usually obtained from a radiosonde equipped with a GPS receiver, such as the one launched each morning during this field course.

The theory behind this technique and an explanation of the calculations is given below.

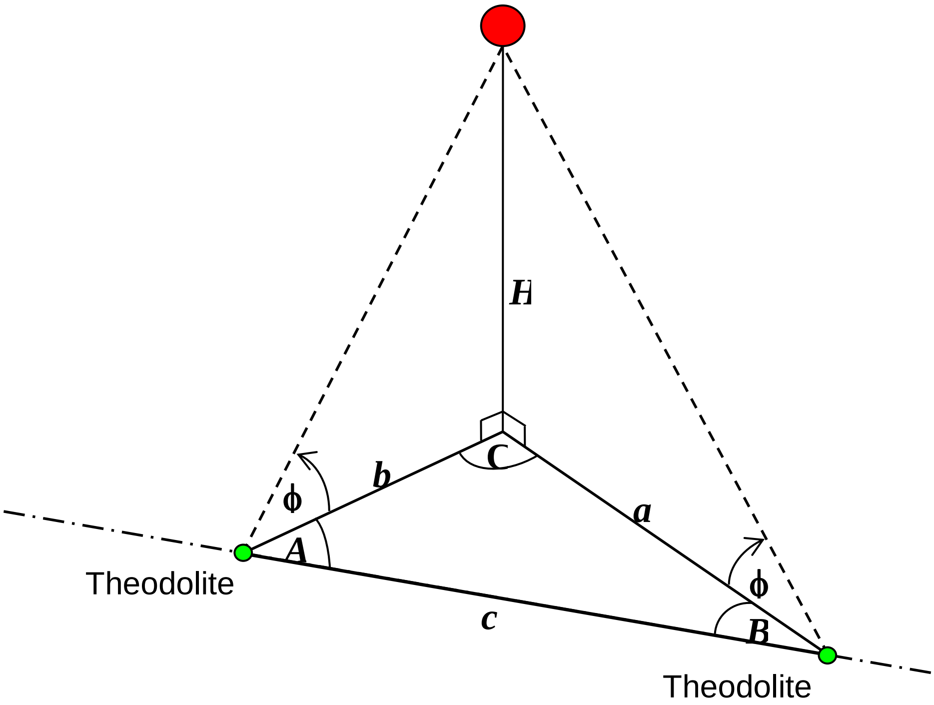


Figure 1. Schematic representation of the tracking of a pilot balloon.

Figure 1 shows the essential features of this method of tracking a pilot balloon. Theodolites are located at positions *A* and *B*; *C* is the position on the ground directly beneath the balloon; these three points define a triangle *ABC* with sides of length *a*, *b*, and *c*. Side *c*, between the theodolites, can be measured directly if the distance is short enough, or calculated from the GPS co-ordinates. You could the great circle distance formula if your GPS co-ordinates are Lat/Long (L_A/G_A and L_B/G_B) and or Pythagoras's theorem if you have OS grid coordinates. The great circle formula is

$$c = R_e \cos^{-1}(\cos L_A \cos L_B \cos(G_A - G_B) + \sin L_A \sin L_B) \qquad (1)$$

where Re is the radius of the Earth (6371000m). In order to determine the altitude of the balloon, *H*, we need to know either the length of side *b* and the angle of elevation ϕ_A**,** or side *a* and ϕ_B. The angles of elevation, ϕ_A and ϕ_B, are measured directly by the theodolites; the lengths of sides *a* and *b* must be determined from *c* and the interior angles of the triangle *ABC* using the trigonometric identity

$$\frac{a}{\sin A} = \frac{b}{\sin B} = \frac{c}{\sin C} \qquad (2)$$

The angles *A* and *B* are not measured directly but can be obtained from the measured angles of azimuth as shown below.

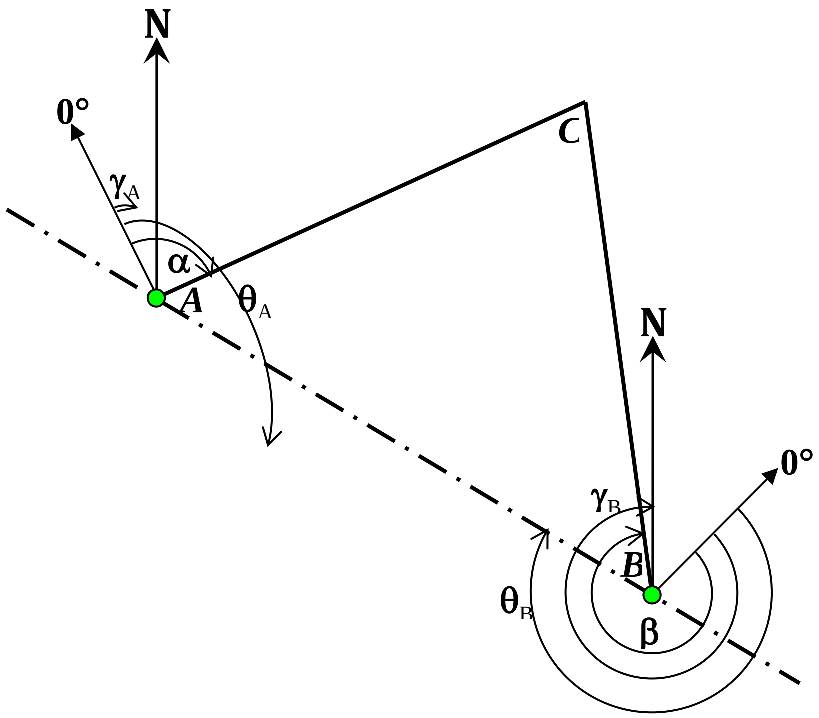


Figure 2. Plan view of the triangle *ABC* for the general case. The directions of true north and of the zeros of the theodolite's azimuth scales are indicated. Note that all angles are measured clockwise (i.e. they are positive clockwise).

Figure 2 shows a plan view of triangle *ABC* for the general case in which no particular care has been taken in orienting the theodolites so that the zero positions of the azimuth scales are set in random directions. In this general case, in order to determine both the interior angles *A* and *B*, and the true bearing of *C* from *A* and *B* it is necessary to measure the angles:

α and β	: the azimuth of *C* from *A* and *B* respectively
θ_A and θ_B	: the azimuth of *B* from *A*, and of *A* from *B*
γ_A and γ_B	: the offset of true north from *A* and *B*

The angles *A*, *B*, and *C* are then given by

$$A = \theta_A - \alpha$$

$$B = \beta - \theta_B \qquad (3)$$

$$C = 180 - (A + B)$$

and the true azimuths from north are given by

$$\text{True azimuth from } A = \alpha - \gamma_A \qquad (4)$$

$$\text{True azimuth from } B = \beta - \gamma_B$$

Once the lengths of sides *a* and *b* have been determined we can calculate *H* from either (or both) of them as

$$H = a\tan(\theta_B) \qquad (5)$$

$$H = b\tan(\theta_A)$$

Having two independent measures of *H* provides a useful check on the accuracy of the measurements used.

The determination of the angles *A* and *B* given in (3) applies only to the configuration shown in figures 1 and 2 – the terms in (3) change with the relative orientation of the zero positions on the theodolites azimuth scales, the orientation of the line *AB* with respect to north, and as the balloon crosses the line through *A* and *B*. In order to simplify the system the theodolites should be aligned in a more careful manner with the azimuth scales zeroed pointing in the same direction along the axis through *A* and *B* as shown in figure 3.

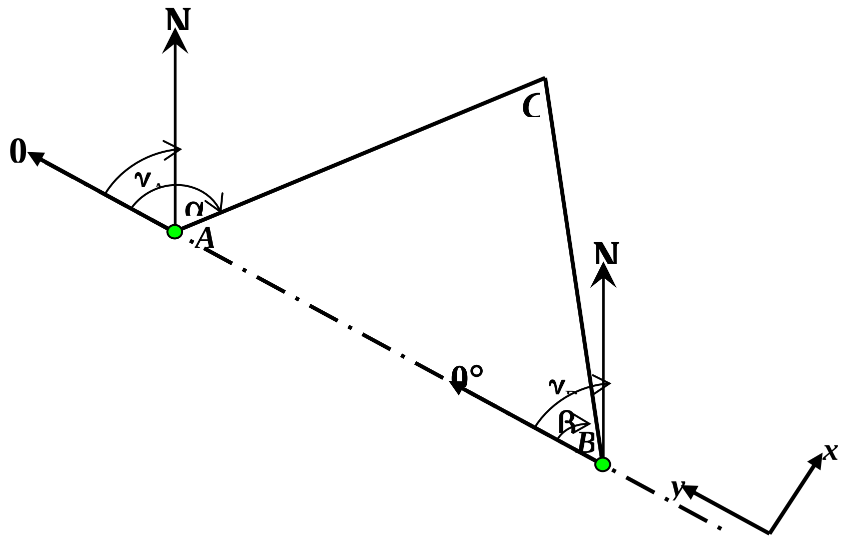


Figure 3. Plan view of the triangle *ABC* where the theodolite's azimuth scales have their zeros aligned with the line *BA*.

In this case $\theta_A = 180°$ and $\theta_B = 0°$. The general case can be transformed to this simpler arrangement by applying offsets to all the measured angles. The offset at $A = 180 - \theta_A$, and the offset at $B = -\theta_B$, thus:

$$\alpha_{new} = \alpha + (180 - \theta_A)$$

$$\beta_{new} = \beta - \theta_B$$

$$\gamma_{A(new)} = \gamma_{B(new)} = \gamma_A + (180 - \theta_A) \tag{6}$$

The software we are using this year does not make this correction, so it is important to align theodolites as shown in figure 3. All the subsequent calculations are carried out in a reference frame aligned with *BA*; the orientation of the *x* and *y* axes is indicated in figure 3.

In order to find the wind speed, we must first find the *x* and *y* coordinates of the point *C*; these are simply given (with respect to *A*) by:

$$x = b\sin\alpha = a\sin\beta$$

$$y = b\cos\alpha = a\cos\beta - c \tag{7}$$

The mean wind speed components for the layer between consecutive measurements are then simply d*x*/dt and d*y*/dt. These are rotated into the Earth based reference frame as follows:

$$u = \frac{dx}{dt}\cos\gamma \quad - \quad \frac{dy}{dt}\sin\gamma$$

$$v = \frac{dx}{dt}\sin\gamma \quad + \quad \frac{dy}{dt}\cos\gamma \tag{8}$$

where *u* is positive to the east, and *v* is positive to north.

The process described above fails when the balloon lies directly over the line through *A* and *B*, when the triangle *ABC* collapses to a straight line. In this case the *x* position coordinate is zero, and y and H must be determined from the triangle in the vertical plane, shown below – this is left as an exercise for the reader.

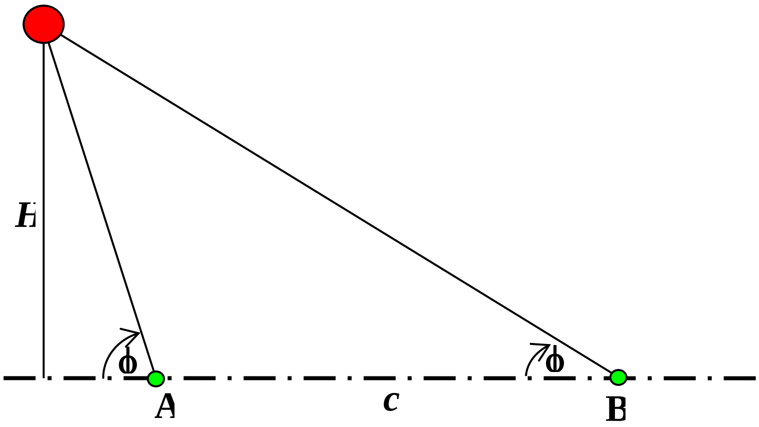


Figure 4. The triangle in the vertical plane, formed when the balloon lies directly over the line through *A* and *B*.

The redundancy noted in equation (5) and the need to use both elevations when the balloon is directly over the line *AB* are not the only problems in finding the balloon's position. Once the balloon is at a distance of more than a few times c, both theodolites are providing almost exactly the same data, and hence the error in our estimate of the radial distance of the balloon becomes very large. This year, we are using an approach which

a) Uses all four angles to estimate the position
b) Uses the previous two locations of the balloon to estimate its current position --- this estimate is then combined with the measurements for the current position. (This approach is known as a Kalman filter.)

<u>Single theodolite method</u>

Lone observers could still estimate a wind profile from a balloon by carefully filling the balloon with a certain amount of gas to give it a known buoyancy, and thereby estimate an average ascent rate, w. Knowing this, height is just a function of time since release:

$$H = wt \tag{9}$$

The distance of the balloon from the observer at A is then given by

$$r_A = H/\tan\phi_A \tag{10}$$

The co-ordinates of the balloon are thus given by

$$x_A = r_A \sin\alpha \quad y_A = r_A \cos\alpha \tag{11}$$

The velocity components u and v are calculated in the usual way using equation (8).

The balloon tracking software produces the following plots:

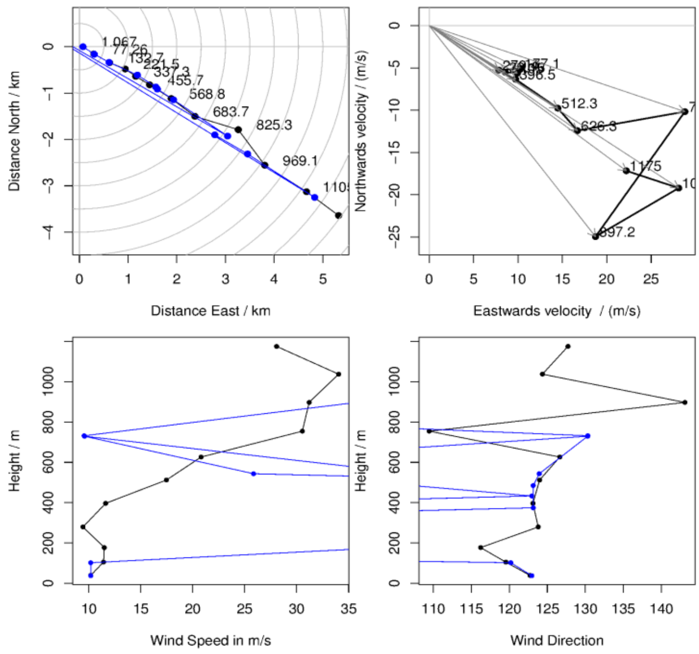


Figure 5. Summary plot. Top left: distance vs time; points labelled with height in metres. Top right: hodograph of wind speeds and directions. Bottom: Wind speed and direction as a function of height. The blue lines are produced using the direct inversion of equations 1-7; the black lines from the Kalman filter.

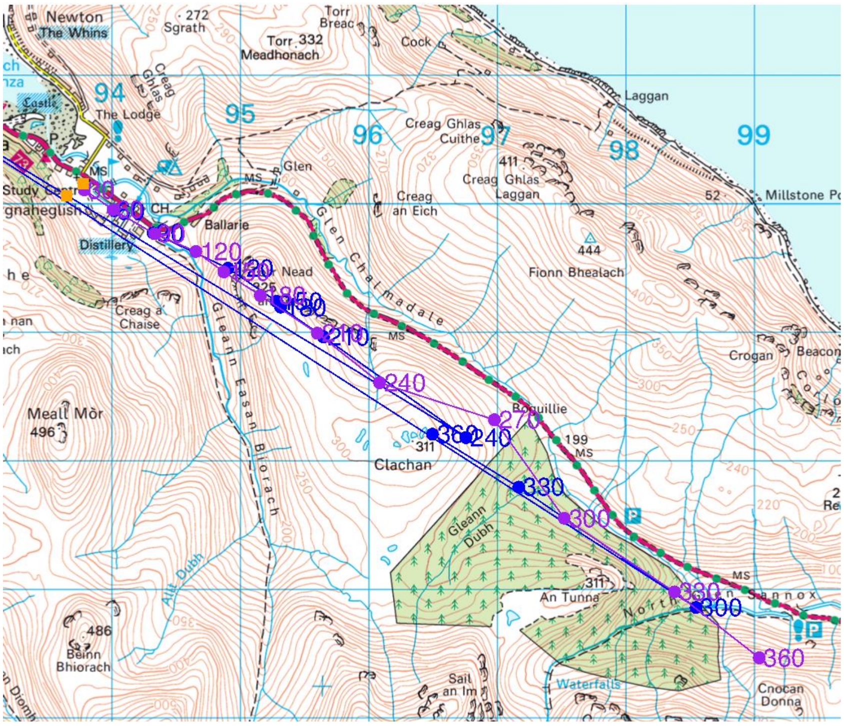


Figure 6. Path of balloon superimposed on OS map. Orange squares mark the two theodolites. Kalman filter results in purple, direct geometric results (equations 1-7) in blue. Points are labelled with time in seconds.

Analysis

NOTE: If for some reason you did not get any convincing profiles you may take data from another group. In this case please put a note in your book explaining this and providing any example data illustrating that your profiles didn't work.

Taking one of your group's balloon profiles:

1. Examine the plot of the trajectory of the balloon obtained from the two methods. Are its starting point and general direction consistent with your observations? If not, check that the columns in your data were in the correct order, and try again. Where do the two methods agree with each other? Where do they disagree? Are there any points which seem unphysical, where the balloon appears to behave inconsistently? [10 marks]

2. Compare the balloon trajectory with synoptic charts and observations available at the same time as the ascent. How does the balloon trajectory compare with 'large-scale' wind? How does it compare with the wind direction/speed observed by the local met mast? [20 marks]

3. Examine the balloon trajectory and the profile of wind speed. What does this tell you about the structure of the flow in the valley? Can you identify any of the flow structures discussed in the background section? [30 marks]

Taking the data from *all* groups:

4. Compare the ascents made at different times and those made by different groups. Do these show the same flow features in the valley? If they do, try to explain why you think this flow pattern consistently occurs in the valley. If not, choose two contrasting examples and discuss why you think they are different. [20 marks]

Errors

Errors in Data

1. **Introduction to Errors** – in the sense of uncertainty in measurement – the understanding of errors are fundamental to scientific exploration. Unless we are counting individual events, the measurements we make will be real numbers with some uncertainty associated with them. To compare our measurements with other measurements or with theory, to combine them with other measurements to get derived products, or simply to publish results, we have to evaluate those uncertainties. That does not mean reducing the errors in every measurement as small as possible - with some ancillary variables a fairly coarse estimate of the error may be all we need, and knowing where to spend time (and money) reducing the errors is one of the skills of an experimental scientist. At Arran we will expect all your measurements to be quoted with an error estimate - how accurate you consider the measurement to be. These errors need not be calculated precisely - they are meant as a guide to uncertainty so one-digit precision is usually enough. Nonetheless errors need careful evaluation in any experiment, and of course experimental design should seek to minimise errors in the final result.

This chapter provides an introduction and reference guide to some of the skills you will need to learn to produce error estimates, and how to combine errors when generating derived products from one or more measurements.

2. **Errors fall into three categories:**

a. Systematic errors. These are errors affecting the whole dataset in the same way. Simple examples are zero and scale (or calibration) errors. Zero errors, where all the measurements suffer from a constant offset, can often be evaluated easily and the data corrected. Calibration errors require the instrument to be checked against a standard - always record with your data when the instruments were last calibrated, and if you don't know, make sure you check the calibration. More complex kinds of systematic error come from instrument response times, drift in sensitivity, or sensitivity to external variables (like temperature or humidity). There is no general theory here - good experimental practice always considers possible systematic errors for each experiment. Always examine your data for evidence of systematic errors, and validate your measurement if you can by comparing with a different technique. Calibration and validation are the key to evaluating and reducing systematic errors.

b. Random errors. These are errors that affect each measurement differently, and in a random fashion. They can either arise from noise in the instrument (e.g. thermal noise in a detector) or from the fluctuations in the quantity being measured (e.g. measuring photon flux at low light levels).

c. Representativity errors. This category of error is very important for fieldwork. How representative is a measurement of (say) temperature measured outside the field centre? How much variation in temperature is there within 100 m? within 1 km? Over an hour? There is no point quoting a measurement to 4 significant figures if its variation in the environment you're working in is ±10%.

3. Random errors.

a. Scale-reading or digitising error. We can read a scale or a digital display to a scale division. Within this bound we cannot expect any one value to be more probable than any other value. The probability distribution function for scale-limited errors is a top-hat function (Figure 1)

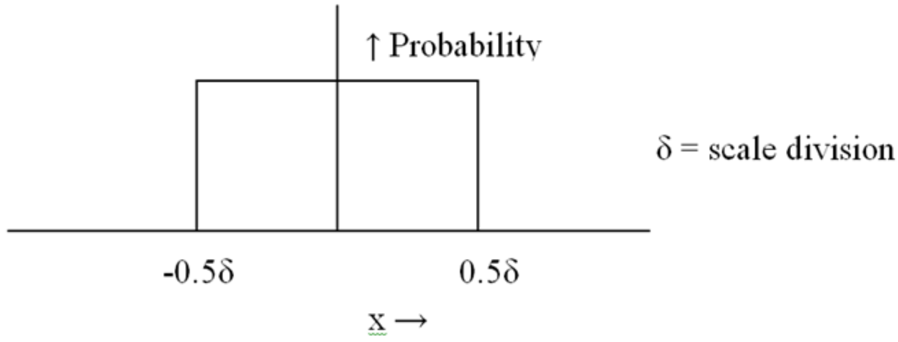


Figure 1: Probability distribution function for scale-limited errors

b. Noise - a stochastic (random) variation in the signal we are trying to measure. We conventionally represent noise as a Gaussian distribution - the mean is the most probable measurement but there is a finite probability of a value well away from the mean.

4. The Gaussian or normal distribution.

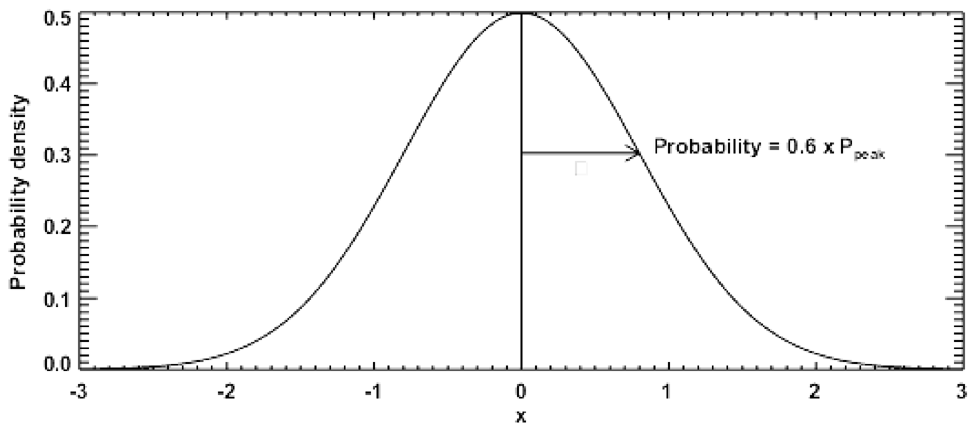


Figure 2: Gaussian distribution with mean zero and standard deviation 0.8. Note that the integral of any probability distribution function (area under the curve) must equal 1.

$$G(x, \mu, \sigma) = \frac{1}{\sigma\sqrt{2\pi}}\, e^{-0.5\left(\frac{x-\mu}{\sigma}\right)^2} \quad (1)$$

where σ is the standard deviation and μ the mean of the distribution. We use G as a model for random (noise-like) errors.

The probability of making a measurement in the range x → x + dx is G(x,μ,σ)dx. It can be shown mathematically that other distributions tend to a Gaussian when there is a large number of contributing events

- distribution of heights in a population of people is Gaussian - the determining factors are genetic and many in number
- distribution of weight in the same population is not Gaussian - it's dominated by one factor, i.e. how much you eat and thus is more of a long-tailed distribution.

As with all probability distributions, $\int_{-\infty}^{\infty} G(x,\mu,\sigma)dx = 1$.

Further, $\int_{x}^{\infty} G(y,\mu,\sigma)dy + \int_{-\infty}^{x} G(y,\mu,\sigma)dy$, which gives the probability of making a measurement more than x from the mean, varies as follows:

x⁄σ	p(|x| > σ)
1	0.32
2	0.05
3	0.003
4	0.0001

i.e. In a sample of values with a Gaussian distribution of random errors, 68% of the values will lie within 1σ of the mean, 95% within 2σ, 99.7% within 3σ and 99.99% within 4σ. 1 in 3 samples should be more than σ away from the mean, 1 in 20 more than 2σ.

5. The Poisson distribution

Although the Gaussian distribution is often assumed to apply in all circumstances, there are certain cases where different distribution functions should be used. An example of a different function is the Poisson distribution, which applies to individual events. For example, suppose the mean arrival rate of photons at a detector is 10 s^{-1}, measured over a suitably long time. Then the distribution of the number of photons measured in successive 1s intervals, r (histogram of no. photons measured) will follow a Poisson distribution:

$$P(r,\lambda) = \frac{e^{-\lambda}\lambda^r}{r!} \qquad (2)$$

where λ is the mean (10 in this case) and r the number of events (in this case a particular value of photons per second, like 8 or 12).

- standard deviation of the Poisson distribution is $\sqrt{\lambda}$
- for λ > 10, the Poisson is well approximated by a Gaussian distribution $G(x,\lambda,\sqrt{\lambda})$.

6. Averaging

If our measurements are limited by noise we can improve our precision by averaging. The **mean** $\bar{x}$ of a sample of N measurements x_i is given by

$$\bar{x} = \frac{1}{N}\sum x_i \quad (3)$$

The corresponding standard deviation is given by:

$$\sigma^2 = \frac{1}{N-1}\sum (x_i - \bar{x})^2 \quad (4)$$

$$= \frac{1}{N-1}\left(\sum {x_i}^2 - N\bar{x}^2\right) \quad (5)$$

(the N-1 comes about because $\bar{x}$ is the mean of the sample not the population mean)

The standard error in the mean - the 1 σ error in $\bar{x}$ - is then:

$$\sigma_m = \frac{\sigma}{\sqrt{N}} \quad (6)$$

This implies that if we take enough measurements we can make σ_m as small as we like. This is true - so long as the underlying signal remains constant, and the measurements remain independent. To understand the latter statement, consider a micrometer being used to measure the thickness of a bar, 0.1 m long. The micrometer head is ~ 0.5 cm wide, so 10 measurements evenly spread along the bar will be independent. If we take 1000 measurements, however, they will overlap, and will no longer be independent.

Although you can average noise-limited signals, you cannot average scale-limited signals to improve precision. If you make 100 measurements of 10.1 cm with a ruler marked in mm, the average measurement is still 10.1 ± 0.05 cm. This leads to an apparent paradox - you can get better precision by adding noise to your signal so that you get a spread of values, then averaging a large number of them! Believe it or not, this is actually done e.g. ‘dithering’ with analogue-to-digital converters!

7. Quoting errors

In each measurement we make there will be some random error, which must be quoted with every measurement. For scale-limited errors we quote half the scale division, but for noise-like error there is a choice - some people use ±1σ, others ±2σ and others again some other multiple of σ. The convention differs in different areas of science and no recommendation either way is given here. But it is very important that when quoting error bars, especially noise-like errors, that you say what they are. Recall that there is a 32% probability that the true value lies outside 1σ of the measurement given, but there is only 5% probability that it lies beyond 2σ.

It is sensible practice to quote random and systematic errors separately, since random errors affect each point differently while systematic errors affect them in the same way. So, for instance, if you wanted to use $z = x_1 - x_2$, with x_1 and x_2 being measured by the same micrometer, a zero error in the micrometer would not affect the difference.

Should we quote errors as absolute or percentage values? Very often the two are equally valid and it is a matter of personal choice which to use. Sometimes, though, it does matter so some care is needed. If we have a set of measurements with the same instrument spanning a range of values, for which the absolute error is the same, it obviously makes more sense to quote absolute error. On the other hand, if we are combining measurements in formulae involving multiplication and division (see below) it is better to use percentage or fractional errors.

8. Some nomenclature

Accuracy and precision. Broadly speaking, the precision of a measurement is its random error. Its accuracy also includes the systematic error. So you can have a very precise, inaccurate measurement but not a very accurate imprecise measurement - accuracy is limited by precision.

Absolute standard error is the σ value of $(x_1, x_2 \ldots, x_N)$

Fractional standard error is $\sigma/\bar{x}$.

Percentage standard error is $100\sigma/\bar{x}$

9. **Combining datasets.** We have already seen how to combine N measurements of a quantity x to find $\bar{x}$ and σ . What if the measurements have error bars?

a. All the error bars are the same. Use the same formulae as before for $\bar{x}$ and σ. But the value of σ (and therefore the error in $\bar{x}$) does not depend on the error bars! This is because there is an implicit assumption that the x_i are distributed in a Gaussian fashion - so the scatter of the x_i gives a measure of their error bar.

b. If all the error bars are different then we do need to use them to calculate $\bar{x}$ and σ. Suppose we have N measurements x_i, each with error e_i, then we write $w_i = 1/e_i^2$ whence

$$\bar{x} = \frac{\sum w_i x_i}{\sum w_i} \quad (7)$$

$$\frac{1}{\sigma^2} = \sum \frac{1}{e_i^2} \quad (8)$$

10. **Combining measurements.** We often have to find the error in a combination of two measurements. Formulae for this are summarised below. First, though, we must emphasise that **all the formulae apply to INDEPENDENT measurements with Gaussian-distributed errors.**

Suppose we measure a, b and c and form $x = ab$, $y = ac$ and $z = xy^2$ .Then x and y are not independent, and we cannot calculate the error in z from the errors in x and y. We must write $z = a^3bc$ and work with the primary measurements.

*** Always express formulae in terms of primary (measured) quantities when you calculate errors.**

Formulae for combining errors in independent quantities - errors add IN QUADRATURE.

How does adding in quadrature work?

If x and y are independent quantities, the error in x is just as likely to (partially) cancel the error in y, as it is to add to it. In fact the correct way of combining independent errors is "to add them in quadrature". In other words, instead of saying (for the case of sums and differences) $d\text{q} = d\text{x} + d\text{y}$, we say $d\text{q} = (d\text{x}^2 + d\text{y}^2)^{1/2}$ (This is what is meant by adding in quadrature). If we think of dx, dy and dq as being the sides of a triangle:

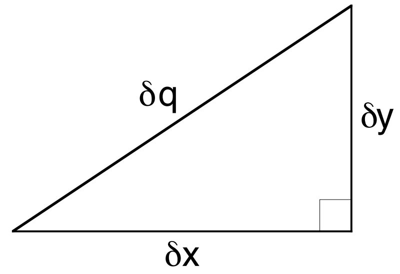


Figure 3: Depiction of Adding Errors in Quadrature

we can see that we have a right-angled triangle, and the rule for adding in quadrature is just Pythagoras' theorem.

Note:

- the error in the result, dq is greater than the errors in either of the measurements, but
- it is always less than the sum of the errors in the measurements.

So the process of adding in quadrature has reduced the error, as we required. It is possible to show rigorously that this is the correct procedure for combining errors.

Formulae for adding errors:

$$z = nx \qquad \sigma_z = |n|\sigma_x \tag{9}$$

$$z = x + y \qquad \sigma_z^2 = \sigma_x^2 + \sigma_y^2 \tag{10}$$

$$z = x - y \qquad \sigma_z^2 = \sigma_x^2 + \sigma_y^2 \tag{11}$$

$$\begin{cases} z &= xy \\ z &= \frac{x}{y} \end{cases} \qquad \left(\frac{\sigma_z}{z}\right)^2 = \left(\frac{\sigma_x}{x}\right)^2 + \left(\frac{\sigma_y}{y}\right)^2 \tag{12}$$

$$z = x^n \qquad \frac{\sigma_z}{z} = |n|\frac{\sigma_x}{x} \qquad (13)$$

For 3 or more quantities the formulae generalise as expected:

$$z = w + x + y \quad \sigma_z^2 = \sigma_w^2 + \sigma_x^2 + \sigma_y^2 \text{ etc.} \quad (14)$$

General cases:

$$z = f(x) \qquad z = f(x_1, x_2, \ldots) \qquad (15)$$

$$\sigma_z = \frac{\partial f}{\partial x}\sigma_x \quad \sigma_z^2 = \sum \left(\frac{\partial f}{\partial x_i}\right)^2 \sigma_i^2 \quad (16)$$

e.g.

$$z = e^x \quad \text{then } \sigma_z = z\sigma_x \text{ i.e. } \sigma_z/z = \sigma_x \qquad (17)$$

$$z = \ln(x) \quad \text{then } \sigma_z = \sigma_x/x \qquad (18)$$

$$z = e^x \tan y \quad \text{then } \sigma_z^2 = (e^x \tan y)^2\sigma_x^2 + (e^x \sec^2 y)^2\sigma_y^2 \quad (19)$$

11. Combining many measurements

Usually the error in one or two of these dominate the final answer so you don't need to grind through a laborious error calculation. One of the key skills of experimental science is to identify the largest source of error in your experiment - and to spend your time (and resources) reducing that rather than fretting about inconsequential errors elsewhere. For example, if we want to evaluate z = xy where the error in x is 3% and the error in y is 1%, then the percentage error in z is $\sqrt{3^2 + 1^2} = \sqrt{10}$ = 3.16%. But we only need the error to one significant figure, so we could have spotted that the error in x is going to dominate the final error, and so avoided the calculation and quoted an error in z of 3%. Knowing when the error formulae are needed and when they are not is a skill that comes with experience but it is rare that one needs a full error calculation involving more than two variables.

12. Some examples of combining measurements

a. Suppose x = 12 ± 3 m and y = 15 ± 3 m, where the errors are 2σ values and x and y are independent.

What is z = x + y? We use Equation 10 to find:

z = 27 m and σ_z = $\sqrt{3^2 + 3^2}$ = 4.2 m.

We quote z = 27 ± 4 m, where again the error is 2σ.

What is z = xy?

Now we must use the fractional or percentage errors, Equation 12

$x = 12$ m ±25% and $y = 15$ m ±20%.

Then the percentage error in z is $\sqrt{25^2 + 20^2}$= 32%. We could quote either z = 180 m^2 ± 30% or z = 180 ± 60 m^2 noting that the errors are 2σ

b. The Rayleigh scattering cross-section for a particle of radius a with radiation of wavelength λ is proportional to $a^6\lambda^{-4}$.

If a = 1.00 ± 0.05 cm and λ = 25 ± 1 cm what is the percentage error in the scattering cross-section?

We first note that the formula involves division (not addition or subtraction) so we are dealing with fractional or percentage errors. We can approach the problem in one of two ways: either evaluate the percentage errors in a^6 and λ^{-4} separately using Equation 13 and combine using Equation 12, or use Equation 16 directly.

Using the first approach, the percentage error in a is 5% so that in a^6 is 30%, and the percentage error in λ is 4% so that in λ^{-4} is 4|- 4|% = 16%. To calculate the percentage error in the cross-section we take $\sqrt{30^2 + 16^2}$ = 34% - or we could argue that the error in a dominates the calculation and quote the final error as 30%. As we are usually only concerned with the first significant figure in the error estimate both are equivalent.

c. Suppose x = 12 ± 3 m and y = 15 ± 3 m again, but this time we want to calculate

$z = \sqrt{x^2 + y^2}$. Again there are two approaches.

Firstly, use Equation 16 directly:

$\partial z/\partial x = x/z$ and $\partial z/\partial y = y/z$, so $\sigma_z^2 = (x^2\sigma_x^2 + y^2\sigma_y^2)/z^2$.

Since $\sigma_x = \sigma_y = \sigma$ for this case, $\sigma_z^2 = \sigma^2(x^2 + y^2)/z^2 = \sigma^2$.

Thus σ_z is the same as σ_x and σ_y in this case!

Alternatively, write $z^2 = x^2 + y^2$ and substitute $a = x^2$, $b = y^2$ and $c = z^2$.

So $c = a + b$ and $\sigma_c^2 = \sigma_a^2 + \sigma_b^2$.

Now from Equation 13, $\sigma_a/a = 2\sigma_x/x$ and so $\sigma_a = 2x\sigma_x$, $\sigma_b = 2y\sigma_y$, $\sigma_c = 2z\sigma_z$.

Writing $\sigma_x = \sigma_y = \sigma$ for this case, $4z^2\sigma_z^2 = 4\sigma^2(x^2 + y^2) = 4\sigma^2z^2$, from which $\sigma_z = \sigma$ as before.

13. **Representativity errors**. These often appear as random errors in the data and if there are repeated measurements they can be treated like random errors. The main difference lies in the interpretation - we need an appreciation of the natural variability in space and time of the quantity being measured (e.g. wind) rather than some property of the instrument (e.g. electrical noise) or of the measurement technique (e.g. photon noise).

Both the independence and Gaussian nature of representativity errors need to be questioned. For example, when measuring a quantity like wind or temperature, we may be able to measure at very high frequency, like 100 Hz. From the instrument's point of view the measurements are independent, but from the atmosphere's point of view they are not. We need to think about correlation structures in the data - how well correlated is a measurement at time t to a measurement at time $t + \tau$? Formally, we can compute the autocorrelation function of the data and examine at what lag this gives a correlation of 0.5 (say) to give us a correlation time. Let's suppose we measure winds for an hour and find that the correlation length is 1 minute. That would imply that we actually only had 60 independent measurements rather than the 360,000 individual values from our 100 Hz probe!

Likewise we need to be sure that we don't have a long-tailed distribution in our variable. An example of this is the aerosol size distribution $N_r(r)dr$ where N(r) is the number of particles in a size range r to r + dr. Atmospheric aerosol range in size from a few nm to a tens of μm, and there are very many more tiny particles than big ones. The distribution is so skewed that a log-normal distribution is usually used to represent the aerosol spectrum. Note however that when calculating the distribution of particles by mass, $N_m(r)dr$, most of the mass comes from the larger particles.

14. **Graphing data.** We generally make measurements to test some theory or other. Let us suppose we vary parameter x and observe the change in y. This may be relevant in fieldwork if x is time or position which we can measure accurately.

By convention, we assume that one parameter has much smaller % errors than the other. The idea is that one parameter is controlled accurately and the other is measured as well as we can. Most formulae you find in computer packages for fitting lines and curves assume that your data are of this form.

We call the controlled variable x and the measured variable y, by which we mean that $\sigma_y/\text{range}(y) >> \sigma_x/\text{range}(x)$. If we expect a linear relation between x and y then we simply plot y vs. x.

What if we expect a non-linear relation, $y = x^2$ say or $y = e^x$? If we want to calculate regression parameters, it is better to transform the equation to y = Af(x) + C and plot y vs. f(x). This should give a straight line whose slope we can measure without changing the errors in y.

Transform x not y - if you transform y the error bars won't be the same!

15. **Fitting a straight line** - the standard method of Linear Regression.

Almost all computer packages will fit a straight line through data but very few will tell you the assumptions they make.

We assume that $\sigma_y/\text{range}(y) >> \sigma_x/\text{range}(x)$ and that all errors in y are equal.

Regression normally uses the least squares principle. Given data points x_i, y_i, i = 1,…N we want to fit a line y = mx + c through the data. So we minimise sum of squares of residuals.

$$S = \sum_{i}^{N} (y_i - mx_i - c)^2 \qquad (20)$$

By taking $\partial S/\partial m = 0$ and $\partial S/\partial c = 0$ we get

$$m = \frac{\sum (y_i - \bar{y})(x_i - \bar{x})}{\sum (x_i - \bar{x^2})} \qquad c = \bar{y} - m\bar{x} \text{ (note that } m \text{ and } c \text{ are not independent)} \qquad (21)$$

$$\frac{\sigma_m^2}{m^2} = \frac{\frac{1}{r^2} - 1}{N - 2} \qquad \sigma_c^2 = \sigma_m^2 \bar{x}^2 \qquad (22)$$

where $r = \frac{\sum (y_i - \bar{y})(x_i - \bar{x})}{(\sum (x_i - \bar{x})^2 \sum (y_i - \bar{y})^2)^{\frac{1}{2}}}$, the correlation coefficient. (23)

16. What if the y_i have different errors? Then we form weights $w_i = \frac{1}{\sigma_i^2}$, and calculate

$$m = \frac{\sum w_i (y_i - \bar{y})(x_i - \bar{x})}{\sum w_i (x_i - \bar{x})^2} \qquad r = \frac{\sum w_i (x_i - \bar{x})(y_i - \bar{y})}{\sqrt{\sum w_i (x_i - \bar{x})^2 \sum w_i (y_i - \bar{y})^2}} \qquad (24)$$

σ_c, c and σ_m are unchanged $(\bar{x^2} = \frac{1}{N} \sum w_i x_i^2)$.

17. What if there are errors in x and y? i.e. *σ_x/range(x) ~ σ_y/range(y)* ?

Basically, if all the x error bars are equal and all the y error bars are equal - so that there is a single σ_x and σ_y (σ_x not necessarily the same as σ_y) then there are formulae for calculating m, c, σ_m, σ_c . But if all the error bars are different there are no such formulae: we have to find a numerical solution.

We meet this kind of situation in atmospheric science when we plot one measured variable against another, looking for tracer-tracer correlations for example. In this case there is no preferred x or y variable and the regression of y on x is as valid as the regression of x on y. We would expect that the two slopes (say m_1 and m_2) would be reciprocals - i.e. $m_1 m_2 = 1$. But this is not what we find! In fact the product of the two slopes is r^2 so if the correlation coefficient is far from 1 the two lines will be quite different. Crucially, a basic line-fitting package will tell you that the error in the slope is much smaller than the difference between m_1 and $1/m_2$.

So, if you're fitting straight lines to data where there are random and representativity errors in x and y, do not believe the error you get from your fitting package! For a rule of thumb that is easy to implement I would suggest taking the best estimate of m to be $0.5(m_1 + 1/m_2)$ and the 1σ error in m to be $0.5|m_1 - 1/m_2|$.

18. Representativity errors and independence. Representativity errors arise from differences in the measured quantity over space and time scales that are smaller than those we are looking to investigate. For example, when comparing data at a point with a model simulation using a 20 km grid box, we need some idea of the variability of the measured quantity within a grid box. Similarly, if we want an hourly measurement of a particular quantity we might measure every minute and take the average to find the desired value. Representativity errors (variations in the measured values) are usually close enough to being Gaussian that the standard deviation is a good measure of the overall variability. But the formula for standard error in the mean, $\sigma/\sqrt{N}$, demands that the points are all independent - and if they are dominated by natural fluctuations they will usually not be. As mentioned above, we could determine a correlation length by calculating the autocorrelation function of the data and finding the lag L at which it falls to 0.5. The number of independent measurements would then be estimated as N/L.

The number of points also comes into the formula for the error in the slope of a regression line, and again you could use the autocorrelation of the data points to find the number of independent points. The more practical approach though is to calculate the two regression lines as described above.

Trajectory Analysis

Trajectory Analysis

Often as atmospheric scientists we are interested in where a parcel of air has come from or will go to. You may be concerned about where the emissions from an industrial plant may end up, or where a particularly polluted air parcel originated from. Trajectory calculations can be used to help you come to some conclusions about these kinds of questions.

Frameworks

There are 2 distinct approaches with which air motion can be viewed.

- ❖ The ***Eulerian*** framework focuses upon a fixed grid of boxes with the wind blowing mass from one box on into the next.
- ❖ The ***Lagrangian*** framework focuses upon individual air-masses or parcels being moved around the atmosphere by the wind. There is no fixed grid.

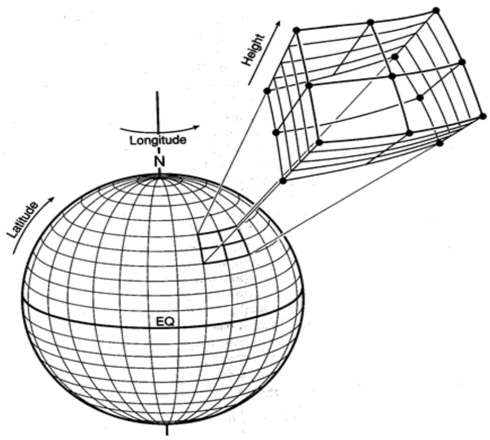


Figure 1: Representation of the Eulerian framework. A fixed grid of boxes is set up and the wind blows mass from one box to the next.

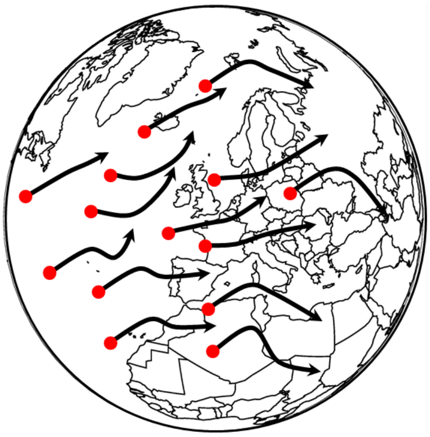


Figure 2: Representation of the Lagrangian framework. There is no fixed grid. Individual air parcels (red dots) are blown around by the winds. The paths mapped out by the air parcels are known as trajectories.

Trajectory analysis is based on the Lagrangian framework. Imagine releasing a neutrally buoyant balloon (stays at the height you put it) into the atmosphere. It is blown around the atmosphere about by the wind. The path that balloon takes as it is blown around is known as the trajectory. These trajectories can be calculated by computer without having to release specific balloons into the atmosphere, and used for a variety of scientific purposes.

How do we calculate a trajectory?

A trajectory is the time integration of the velocity of a parcel of air as it is transported by the wind to provide a time dependent location. This can either occur **forwards** in time, showing you where an air-mass will go, or **backwards** in time, showing you where the air-mass came from. The parcel's transport by the wind is computed from integrating the velocity field either forwards or backwards in time. If $\mathbf{x}(0)$ is the initial position, $\mathbf{x}(t)$ is the position of the parcel at a time t later, and $\mathbf{v}(\mathbf{x},t)$ is the wind velocity then

$$\mathbf{x}(\mathrm{t}) = \mathbf{x}(0) + \int \mathbf{v}\,dt$$

The wind field vector $\mathbf{v}(\mathbf{x},t)$ varies with longitude, latitude, altitude and time (t). These wind fields are normally provided by the national meteorological services (e.g. UK Met Office, European Centre for Medium Range Weather Forecasting, etc). They are a grid of information describing the wind speed and direction over the globe or a smaller region. If you want to make predictions into the future they are based on weather forecasts. If you want to investigate events from the past they are based on 'analyzed winds'. These analyzed winds represent the 'best guess' of the meteorological services of the atmosphere at anytime. They come from a combination of direct observations of the wind direction and speed combined with our knowledge of the physics of the atmosphere. The wind fields give the mean value of each of the 3 components of the wind over a certain timescale. For global scales this might be every six hours, for regional scales every hour. Global wind fields give the mean wind direction and speed every 100 km or so, whereas regional models have a finer resolution at 10 km or so.

If all three components of the wind field provided by the meteorological services are used the trajectories are described as *kinematic*. If the vertical component of the wind in unavailable various approximations can be made.

1) *Constant Altitude*

   The air parcel stays at a constant height above the ground

2) *Constant Pressure (**isobaric**)*

   The air parcel stays at a constant pressure

3) *Constant potential temperature (**isentropic**)*

   The air parcels stays at a constant potential temperature (□h)

4) *From the divergence of the horizontal winds (**kinematic**)*

   If the horizontal winds do not balance each other (ie mass is being pushed into a grid box) the vertical velocity of the wind has to change to balance the flow. You can calculate the vertical velocity from the imbalance in the horizontal wind field. This uses a vertical wind so is a *kinematic* trajectory but this is less good than using the vertical winds provided by the meteorological services.

The integration can be achieved numerically through a variety of standard techniques such as the Runge-Kutta method.

Limiting issues with trajectories

Although trajectories are a useful diagnostic tool they are not perfect. They suffer from a variety of issues that you must be aware of. These issues can be separated into:

Sub-grid issues

Chaos

Accuracy

Sub-grid issues

The wind fields from the meteorological services are averages over the scale of the grid-box. These grid boxes are of the order of 100 km in the horizontal for global models and 10 km for regional model. However, some important atmospheric processes occur on much smaller scales. Clouds and convection occur with characteristic scales on the order of a kilometer. Boundary layer mixing occurs at scales on the order of 500m.

Thus the average velocities for each grid box will 'average out' these scales of mixing and miss the impact of clouds and boundary layer mixing. In some situations this is not a serious issue.

If you are interested in the flow of pollutants across the Atlantic at 5 km, mixing within the boundary layer is not important. If you are interested are in a region without clouds then the inability to represent clouds is not an issue. However, for many applications you need to be aware of the possibility of trajectories not being the appropriate tool for analyzing observations.

Chaos

As an air parcel moves further and further from its origin the chances of small errors in the initial position causing large errors in the calculated position increases. Trajectories started only very small distances apart can rapidly diverge and end up at very different parts of the globe. Thus on some timescale trajectories become inaccurate and cannot provide useful diagnostic information. If you consider a large number of trajectories over a large period of time the rate at which they diverge may be well defined; however it is not for an individual trajectory. The divergence is dominated by small regions of space, thus a set of trajectories may be separated by only a small distance for long periods of time and then suddenly diverge. If these divergence points can be identified they can provide information about the length of trajectory that can be considered appropriate.

Accuracy of the wind field

The quality of the trajectories will only be as good as the quality of the wind fields used as input. As discussed earlier these wind field represent the ‘best guess’ at the state of the atmosphere by the meteorological service. It is a balance between the available observations and the quality of the meteorological model used. Over some regions (e.g. the middle of the Pacific) we have very few observations of what the atmosphere is doing. The analyzed or forecast wind fields provided by the meteorological service will be dominated by what the model thinks should be happening, rather than any constraints from the observations. In this region there may be far larger errors in the wind field than over say Western Europe where a large number of observations are made every day that can be included in the wind fields.

Examples of the use of trajectories

Trajectories have been calculated for a number of years using a variety of methods. Computers are now extensively used for this purpose. One of the initial uses was for the prediction of dispersion from single sources such as industrial incidents at chemical and nuclear plants and also the dispersion of radioactive fallout from the detonation of a nuclear weapon. It is possible to calculate both forward and backward trajectories, both of which have a number of uses including synoptic meteorology investigations such as air-mass flow around mountains, identification of pathways of water vapour or desert dust and environmental applications such as source-receptor relationship studies of air pollutants. Trajectories have even been used to detect illegal marihuana cultivation through analysis of pollen content in air.

Chernobyl.

The Chernobyl accident in 1986 was one of the first well documented examples of the use of trajectory analysis. It was used to identify the route taken by the fallout plume following its emission at the accident site in Belarus in the former Soviet Union. These were forward trajectories, i.e. the source location was known and by using the wind field data the path taken by that air parcel was tracked. Figure 3 shows the trajectories of the air parcels in 2 day increments following the accident, it is clear that the cyclonic system present over the south west approaches on May 3rd drew air off the European mainland over the north east of Britain and the plume hit the ground in particular in the upland areas of Cumbria. The colour indicates the concentration of cæsium 137, a highly active radioisotope, In fact so significant was the quantity of the fallout 'landed' on upland areas of northern England such as Cumbria that local lamb was not marketed to the public for a number of years.

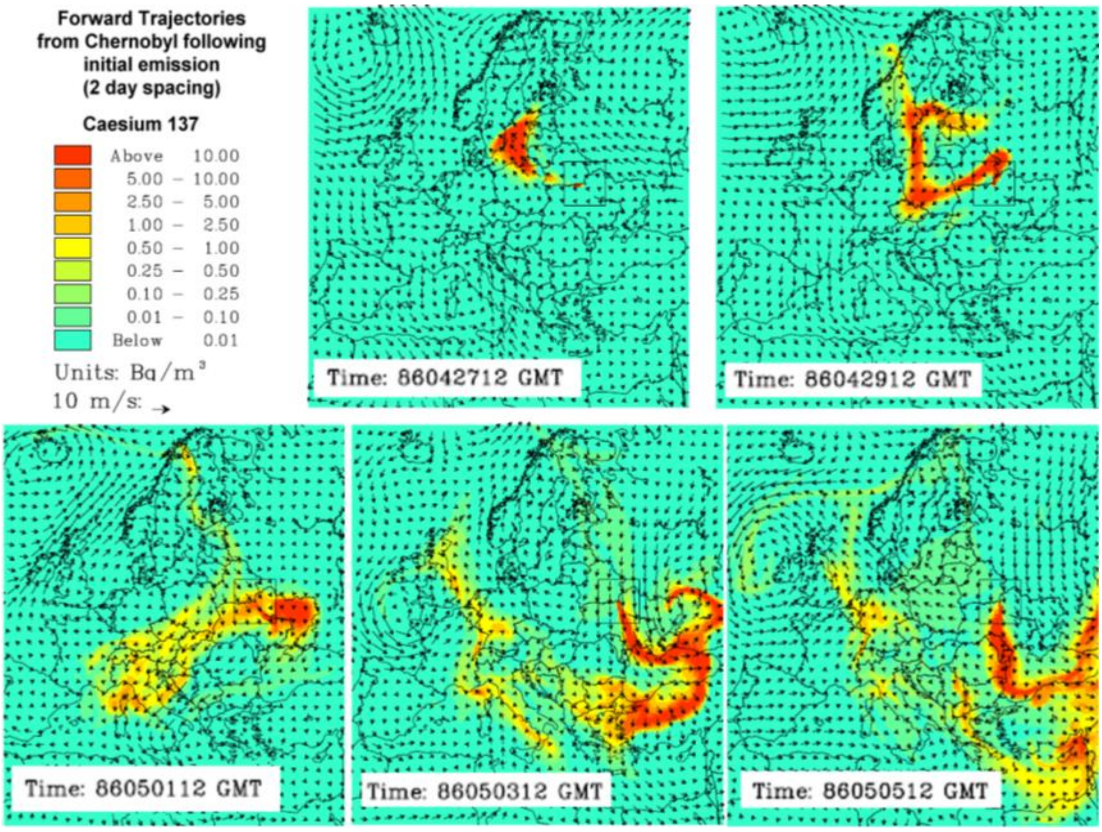


Figure 3. Forward trajectories from the explosion at Chernobyl. Plots show the density of trajectories released from the site every other day at noon GMT. Initially the radiation is confined to eastern Europe however over time large swaths of Europe are influenced by the radiative plume.

Understanding pollution events

Observations of atmospheric composition need to be placed into a meteorological context. Back trajectories can be used for this.

Figure 4 shows data collected on the west coast of Ireland during a research field campaign in 1997. The bottom panel shows the observation of toluene and benzene, two hydrocarbons emitted by industrial activity. It is obvious that there is a large degree of variation in these species. Trajectories were calculated arriving at this site every hour of the campaign. They were then classified into six different types based upon the origin of the air. Typical examples of these types are shown in the upper panels. The trajectories calculated to have passed over industrial regions (France and the UK) show much higher levels of pollutants than those in the other categories.

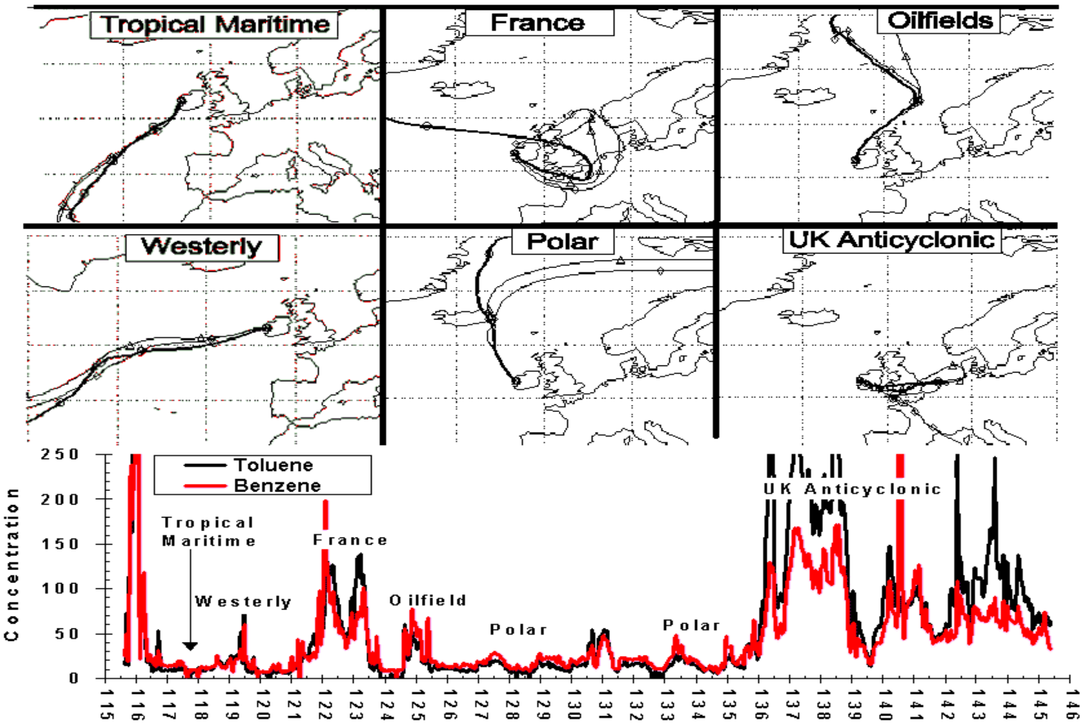


Figure 4. Bottom panel shows the time series of two hydrocarbons measured on the west coast of Ireland in the summer of 1997. High concentrations correlate to air that has passed over France or the UK before arriving at the site. Typical example of trajectories of different origins are shown in the upper panels.

If you compare the relative concentrations of benzene and toluene in air arriving from France and air arriving from the UK there appears to be much higher relative concentrations of toluene in the UK air than in the French air. This reflects the chemical processing of the air. Toluene reacts much faster in the atmosphere with the hydroxyl radical (the main chemical sink for hydrocarbons) so it lost from an air-mass faster. The longer the time between the air being polluted and it arriving at the site the more toluene will have been removed compared to that benzene.

The use of back trajectory analysis has provided extensive insights into the meteorological sources of variability in the atmosphere.

Use of trajectories in Arran.

During the field course you will be making a variety of analyses and observations. Consider when trajectories will be able to provide extra insights into the observations you have made.

To calculate trajectories we will be using the HYbrid Single Particle Lagrangian Integrated Trajectory Model developed by the Air Resources Lab at NOAA (http://www.arl.noaa.gov/ready/hysplit4.html). This is a powerful on-line model which can be run through a web interface.

To use the HYSPLIT model you will need to know the starting (if you are interested in air will go to) or ending (if you are interested in where air came from) point. This should be as a latitude and longitude. **Lochranza is ~55.7N 5.28W**

Click on next and you can now enter more information about the trajectory you want to calculate, such as

Trajectory direction: Backwards or forwards depending on the nature of the problem you are studying.

Vertical motion: Do you want to use the modelled vertical velocity or assume a constant pressure for your trajectory.

Start time: What time do you want the trajectory to start at?

Total run time**:** How for forwards or backwards do you want the trajectory to go?

Start height: How high in the atmosphere do you want your trajectory to start? You can choose 3 levels to start at. AGL is Above Ground Level, AMSL is Above Mean Sea Level.

Plot projection: What map projection would you like to look at the trajectory with.

Vertical plot height units: What would you like the vertical coordinate to be?

Once you have decided press the request trajectory button and wait for the computer in Maryland to calculate the trajectory.

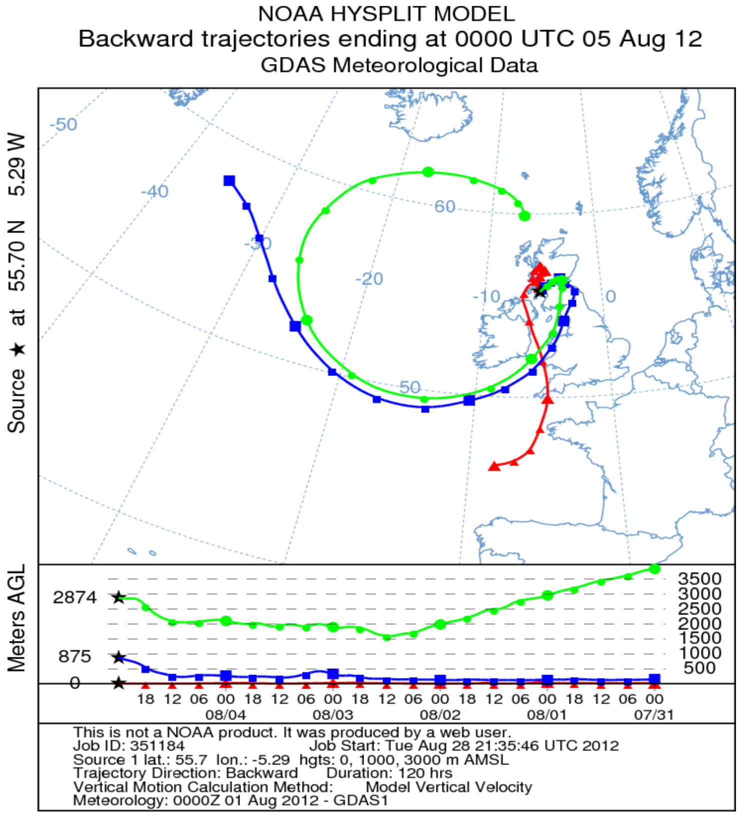
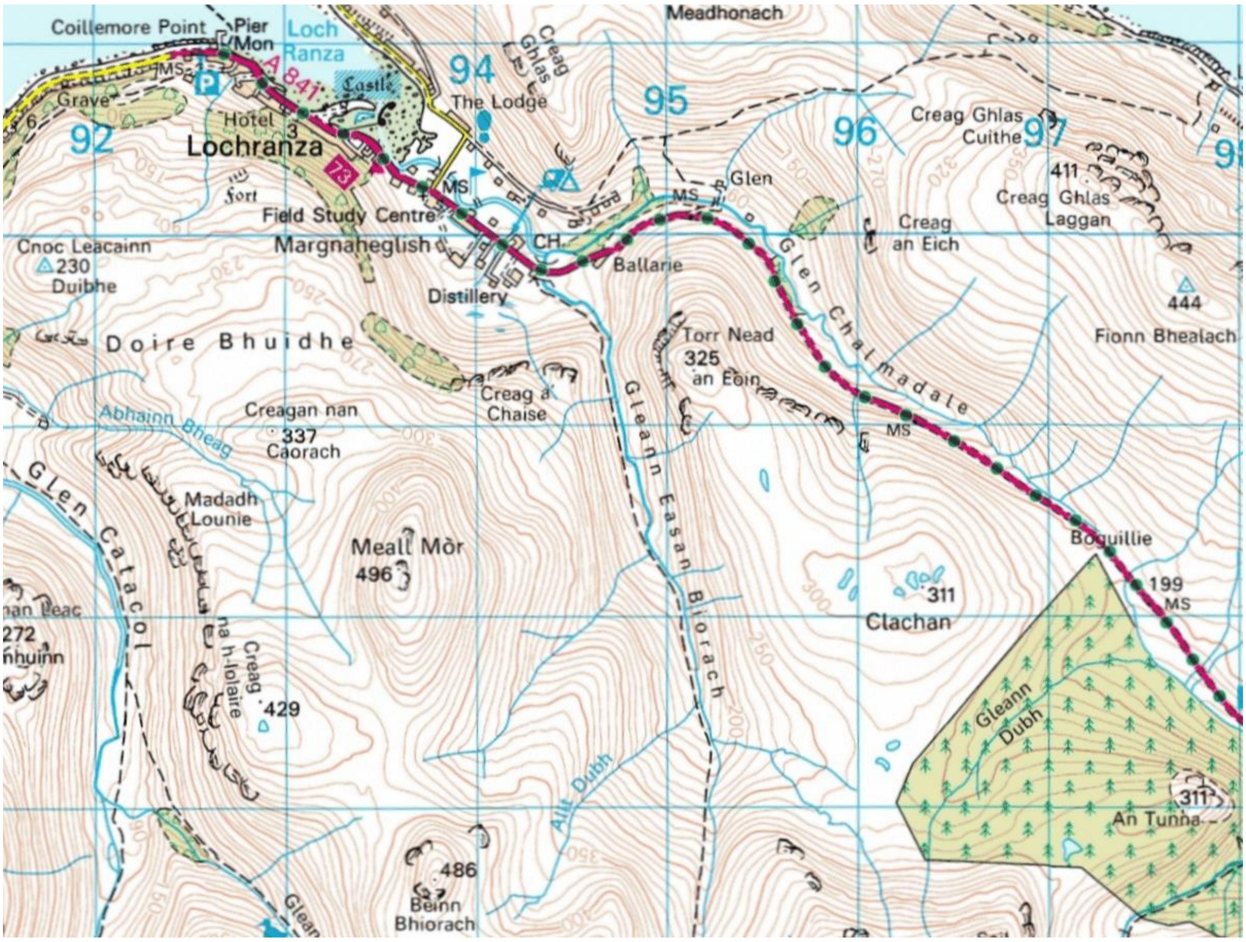


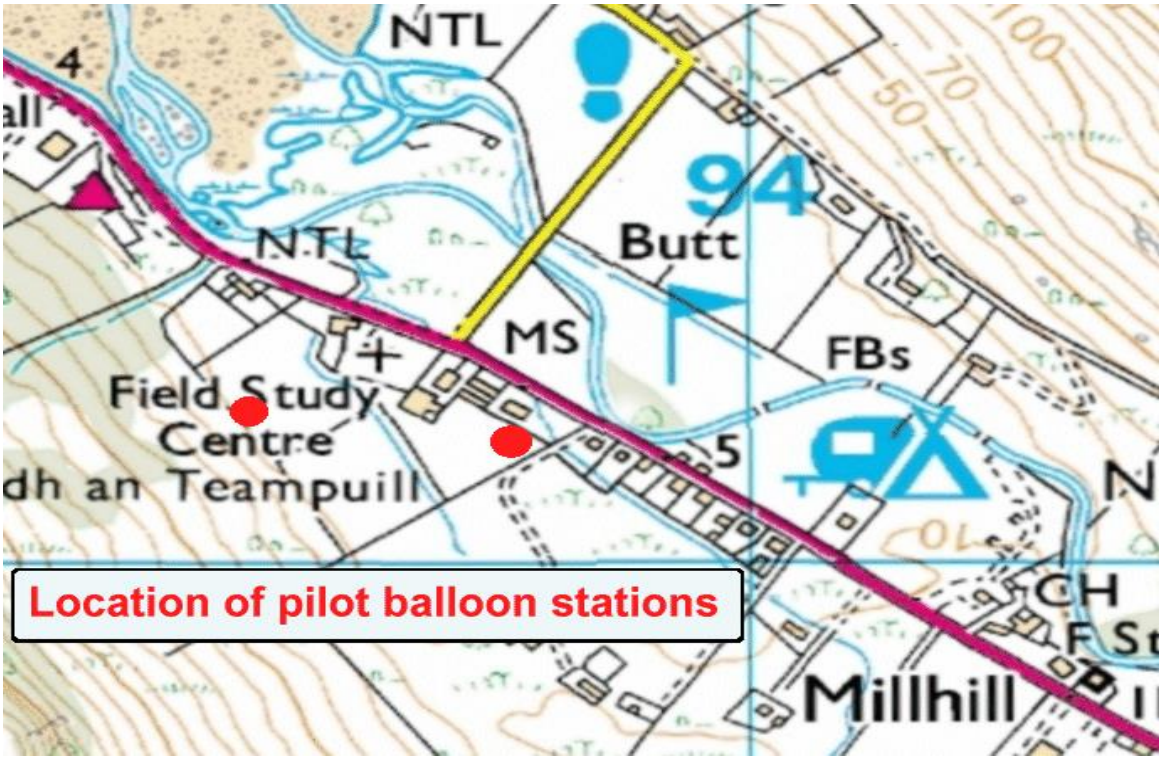
Figure 5: Example of 120 hour, 3 level back trajectory (model vertical velocity), it clearly shows descending air in the free troposphere.

Maps of Lochranza Field Centre

Latitude: 55.70°N Longitude: 5.28°W

The Lochranza site





Risk Assessments

Fieldwork Project Details

Activity: **Ascent of Goatfell, Arran (874 m AMSL) as part of Atmospheric Fieldcourse, Saturday 14th September 2019**

Location:

Goatfell, Arran Island (summit - OSGB NR 991 415).

Brief description of Fieldwork activity and purpose

A relatively strenuous walk, approx 5 hrs in length. This activity is designed to make an ascent from sea level to ~900 m ASL to allow students to gather vertical profiles of a number of basic meteorological observations, including temperature, static pressure, relative humidity.

Travel to and from car park at base of the hill via local bus company (professionally driven coaches). Course has been running since 2003 years and several of the staff have done this route on numerous occasions before.

Organiser Details	***Contact details*** *Name, email, telephone*
Course Leader	***Dr Jim McQuaid, j.b.mcquaid@leeds.ac.uk, Univ phone: 07854400478***
Nature of visit: 10 km walk	Group of approx. 40 UG/masters students + approx 8 academic/research staff; day trip only.

Hazard Identification

Identify all hazards specific to fieldwork trip and activities, describe existing control measures and identify any further measures required.

HAZARD(S)	**CONTROL MEASURES** ***(e.g. alternative work methods, training, supervision, protective equipment)***
Nature of the site Main hazards in grounds are uneven & loose gravel paths, steep slopes, some cobbles & undulating terrain.	The party will walk together following a well trodden route. Reasonable mobile coverage along most of route; emergency services available using normal process. Staff student ratio of at least 1:6 (including qualified emergency first aiders who have completed the 4-day course (Jim McQuaid). Trip participants will be advised to wear suitable footwear and bring appropriate clothing for the weather conditions (waterproofs if wet, appropriate sunhats/suncream if sunny and hot). **Non-compliance will result in exclusion from the walk.**
Process - Trip participants will walk the route above & make atmospheric observations en-route	Most of the staff present have done the walk before and are well versed in the hazards present

Transport – Commercial Coach company (University approved)	Travel to and from car park at base of the hill via local bus company (professionally driven coaches). Evacuation vehicle will be parked in car park at start/end point of the walk, keys left in location known to staff.
Equipment	Meteorological sampling equipment, minimal associated hazard. Temperature probe has a blunt tip, carry in a secure location to avoid puncture injury.
Violence	N/a, very low hazard - rural environment, a walk along very well trodden public footpaths

Individual(s)	Students/staff requiring medication etc will be requested to bring this with them and to notify the trip organisers of their condition prior to departure. Students will also complete a medical information form as part of the induction process.
Work Pattern	The walk will happen in a single day and end well before dusk.
Other- Ticks (carriers of lyme's disease. Arran is well known for having ticks.	• Keep skin covered when in "brushy" areas. • Wear trousers rather than shorts or skirt. • Light coloured clothing makes ticks easier to spot. • Button up collars and cuffs, wear a hat, tuck trousers into socks. • Apply insect repellent on clothes or exposed skin. ***If a tick is discovered contact centre staff who are trained in the removal.*** ***All students and staff are briefed on the hazards from ticks upon arrival at the field centre.***

Additional Control Measures

Training: *Identify level and extent of information; instruction and training required consider experience of workers*

No further training required. 5 hours of relatively strenuous hill walking, with a number of stops along the way for measurements.

Supervision: *Identify level of supervision required*

Students under staff supervision at all times.

Staff will carry mobile phones in order to communicate with each other in case of emergency

Other Controls- *e.g. background checks for site visits*

Most of the staff present have done the several times walk before.

Students will remain with the group at all times during the walk up Goat Fell, staff members will be equipped with first aid kits, survival bags and extra clothing.

Emergency contact numbers: Jim McQuaid – 07751 160953
Hospital: Located at Lamlash, Arran War Memorial Hospital has an A&E department and is well signed locally. Tel 01770 600777
Doctors: The Medical Centre at Lamlash Tel: 01770 600516 /7
• Brodick Medical Centre Tel: 01770 302175
• Inglewood Medical Centre at Shiskine Tel: 01770 860247
Police: Crime on the island is very low, but if you do need to contact the Police for other than a 999 call then there is a station at Lamlash (not always manned) Tel: 01770 302573/4

Identify Persons at Risk

This may include more individuals than the fieldwork participants e.g. other employees of partner organisations	University of Reading/Edinburgh staff and students are working with the Leeds cohort, they follow H&S rules as laid down within this document. They are regular participants in the Arran fieldcourse so well aware of associated hazards.

Additional Information	
relevant to the one working activity including existing control measures; information instruction and training received, supervision, security, increased lighting, emergency procedures, first aid provision etc.	Students will remain with the group at all times during the walk. Staff members will be equipped with first aid kits, survival bags and extra clothing.

Assessment carried out by	Name:	Dr Jim McQuaid
	Signature:	
	Date:	29th August 2019
Names of person(s) involved in Fieldwork	Name:	**A single register will be collected and kept by the module leader, Dr Jim McQuaid**
	Signature:	

General guidance for Arran fieldcourse in addition to that specifically detailed in Goat Fell Risk Assessment.

Hazard	**High**	**Risk** **Med**	**Low**	**Control Measures To Minimise Risk**
Physical Trip hazards		✓		Work is carried out in grassed field – suitable footwear should be worn at all times, waterproof clothing if wet. Tripping on uneven surface or over ground anchors/guy lines/data cables in long grass most notable hazard. Equipment operation is automatic, student will only be inspecting data loggers. During setup of mast, hazards include minor injuries sustained from use of hand tools, and the risk of falling objects during lifting of mast.
Biological Faecal matter (sheep/deer)			✓	Avoid wherever possible, wash hands after working in the field
Ticks (carriers of lyme's disease. Arran is well known for having ticks.		✓		Keep skin covered, Wear trousers rather than shorts or skirt. Button up collars and cuffs, wear a hat, tuck trousers into socks. Wear boots, not sandals. Light coloured clothing makes ticks easier to spot. Apply insect repellent on clothes or exposed skin. ***If a tick is discovered contact centre staff who are trained in the removal.***
Chemical Latex allergy			✓	Helium balloons are made from Latex (no alternative available), staff and students informed of this.
Man Made Falling objects		✓		Students briefed on risks, work only during daylight, and under staff supervision during setup and teardown of equipment.
Road traffic			✓	Apply usual common sense rules when crossing roads.

FOLLOW THE INSTRUCTIONS OF STAFF AT ALL TIMES.

Obey the perimeter notices when the mast is being erected/dropped.

Setup is conducted primarily be experienced members of staff, with all assistance from other staff or students being closely supervised, and all work being inspected. During lifting of mast, all personnel not actively involved must stay outside the potential fall-zone of the mast. Trained personnel assisting during the lifting of the mast will be assigned specific roles, and given instruction on carrying them out.

Other activities

During the fieldcourse personnel fill latex rubber balloons with helium gas, staff dispensing the helium are trained in its delivery and there is full CoSHH documentation available with the cylinder. Students watch the process, a student may be requested to hold the balloon during filling. The gas pressure in the delivery tube is controlled by a standard pressure regulator and flows freely, the pressure in the balloon during inflation is little over local atmospheric pressure.

When sufficient helium has transferred, the balloon is sealed with cable ties and the radiosondes package attached.

In accordance with CAA legal requirements, there is a NOTAM in place for balloon launches at the site daily at 07:00, 13:00 and 19:00 (local time) and also 20g pilot balloons (no payload), between 09:00 and 18:00 (local time).

Detailed risk assessments are provided for staff involved in the specific activities, such as manual handling, tower erection, balloon handling (gas cylinders) and CoSHH for helium gas.

Display Screen Equipment.

During the course, staff and students will be using computing equipment but for relatively short periods of time due to design of timetable.

Non-teaching periods.

Staff are responsible for students during periods during which their performance/work is assessed, outside of these times (~2130 – 0730) the students are permitted free time.

Please do not forget that during the entire stay on the island you are representing your university.