Dynamical influences on European climate: An uncertain future

By Tim Woollings

Climate science is coming under increasing pressure to deliver projections of future climate change at spatial scales as small as a few kilometres for use in impacts studies. But is our understanding and modelling of the climate system advanced enough to offer such predictions? Here we focus on the Atlantic-European sector, and on the effects of greenhouse gas forcing on the atmospheric and, to a lesser extent, oceanic circulations. We review the dynamical processes which shape European climate and then consider how each of these leads to uncertainty in the future climate. European climate is unique in many regards, and as such it poses a unique challenge for climate prediction. Future European climate must be considered particularly uncertain because a) the spread between the predictions of current climate models is still considerable, and b) Europe is particularly strongly affected by several processes which are known to be poorly represented in current models.

Keywords: Jet stream, Storm track, Blocking, Meridional overturning circulation, Stratosphere

1. Introduction

Climate science is coming under increasing pressure to provide projections of possible climate change on regional or even local scales. For example, the UK Department for Environment, Food and Rural Affairs (DEFRA) recently commissioned a consortium led by the Met Office Hadley Centre to provide climate projections for use by the impacts and planning communities. The resulting product (UKCP09; http://ukcp09.defra.gov.uk/) provides users with probabilistic projections of UK climate over the coming century on spatial scales as small as 5km. The consortium have gone to great lengths to account for many different kinds of uncertainty. But do we really have enough confidence in our understanding and modelling of the climate system that predictions on the scale of a few kilometres are justified?

Unfortunately for DEFRA, Europe is arguably one of the hardest regions outside the tropics† for which to predict weather and climate on all timescales longer than a few days. European seasonal forecasts, for example, often show very little skill (*e.g.* Palmer *et al.* 2008), and future climate projections have a particularly large spread between models and a low signal to noise ratio over Europe compared to other mid-latitude regions (Hawkins and Sutton 2009).

There are several reasons why European climate is particularly hard to predict. In general these are related to the large-scale circulations of the atmosphere and ocean which influence European climate. The aim of this paper is to review some of the key processes which shape European climate and to consider how confident we are in the representation of these processes in the climate models used to predict future change. The focus is on dynamical features such as the jet streams and the storm tracks, and on how these features respond to an increase in greenhouse gases. This naturally also leads to a focus on Northern Europe. Projections for Southern Europe at least show stronger agreement between climate models, for example in the projected drying of the Mediterranean (Christensen *et al.* 2007), so appear to have lower uncertainty, at least in the sign of the response. Dynamical influences are strongest in winter, but much of the discussion is applicable to the other seasons, with the possible exception of the influence of the stratosphere. There are, of course, other sources of uncertainty such as the choice of emissions scenario and the effects of aerosols and carbon cycle and cloud feedbacks, but these are less specific to Europe. Local processes, such as the effects of soil moisture and snow cover are also important but are not discussed here.

Natural variability is an important source of uncertainty for the next few decades, and is particularly large over Europe (Hawkins and Sutton 2009). There is hope that some of this variability may be predictable, in particular that arising from variations of the Atlantic Multidecadal Oscillation (AMO; *e.g.* Knight *et al.* 2005).

† Many aspects of tropical climate are also particularly hard to predict for several reasons, such as the importance of cloud feedacks and small-scale convection (*e.g.* Rind 2008).

The European impact of AMO variations is largest in summer, but model results suggest that even then the influence is relatively small (Sutton and Hodson 2005).

2. Current European Climate

In this section we briefly describe some of the key dynamical features which shape European climate. Some of these clearly influence other regions of the globe as well, though we will try to emphasise the factors which help to make European climate unique. While the section is split into subsections, it will become clear that several of these are components of the climate system which interact strongly with each other.

(*a*) *The jet streams*

Figure 1 summarises several aspects of Northern Hemisphere climate. The left-most panels show the winds at 250 hPa†. The westerly (by convention, from west to east) jet streams are seen clearly as regions of high wind speeds in the mid-latitudes. There are essentially two processes which give rise to the westerly jet streams. The first is the poleward momentum/vorticity transport associated with the upper branch of the Hadley circulation‡. Jet streams arising from this process are called subtropical jets. Secondly, the transient eddies which are pervasive in the troposphere lead to vorticity and heat transports which also have the net effect of accelerating the westerly wind, at lower levels in particular (*e.g.* Hoskins *et al.* 1983). Jet streams arising from this are referred to as eddy-driven jets, and in contrast to subtropical jets they extend down through the depth of the troposphere. Both types of jet ultimately owe their existence to the meridional contrasts in solar heating, which provide energy for both the Hadley cell in the tropics and the midlatitude transient eddies. The jet streams are much weaker in summer than in winter, and are generally located closer to the pole. This is because the equator to pole temperature gradient is weaker, and also displaced north, in summer.

† Roughly 10km up in the atmosphere near the tropopause, which is the top of the very active region called the troposphere. Pressure is widely used as a vertical coordinate in meteorology.

‡ The Hadley circulation consists of a bouyancy-driven overturning circulation cell in the tropics, with warm air rising near the equator and moving polewards in the upper troposphere, before sinking and returning equatorward.

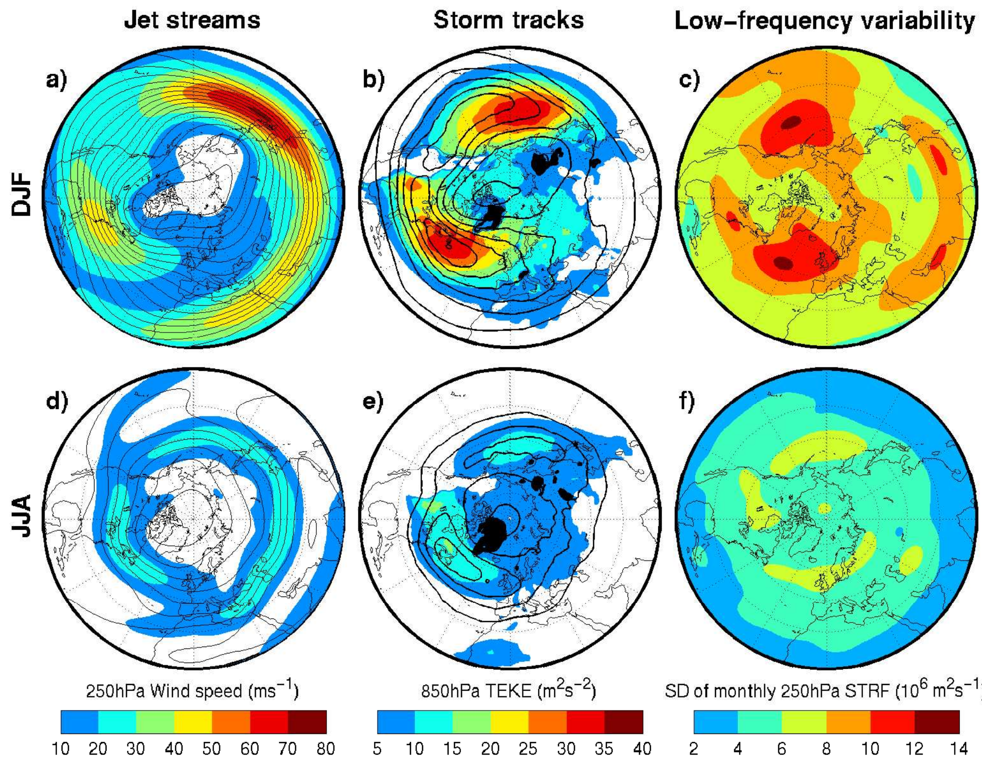

Figure 1. Northern Hemisphere climatology for December 1957 to August 2002 from the ERA-40 reanalysis (Uppala *et al.* 2005). a) December - February (DJF) 250 hPa wind speed with streamfunction contoured every $1 \times 10^7 m^2 s^{-1}$. b) DJF Transient eddy kinetic energy (TEKE) using 2-6 day bandpass filtered winds. Shading shows low level values (850 hPa) and contour lines show the upper level (250 hPa) values contoured every 20 $m^2 s^{-2}$. c) Standard deviation of monthly-mean 250 hPa streamfunction in DJF. d) - f) are as a) - c) but for June - August (JJA).

In many instances there is no clear separation between the subtropical and eddy-driven components of the flow, but the North Atlantic sector is an exception to this. In winter there are two clearly separated jet streams: the subtropical jet at the latitude of northern Africa, and the eddy-driven jet further north. The only other region to show such a clear split between the jets is the South Pacific in the vicinity of New Zealand in the southern winter. The separation of the two jets over the Atlantic is one of the underlying reasons why European climate is somewhat unique.

(*b*) *The storm tracks*

Midlatitude cyclones and anticyclones occur preferentially in certain regions, which are termed the storm tracks. The climate of Europe is strongly influenced by the storms which track across the Atlantic from the east coast of North America, and these make up the North Atlantic storm track. In addition to their obvious societal impacts, the storm tracks are very important components of the global climate system. Again, the storm tracks owe their existence to the meridional contrast in solar heating which drives horizontal temperature gradients. As a result, the troposphere is baroclinically unstable†, and the cyclones and anticyclones grow on this instability, converting available potential energy into eddy kinetic energy. These weather systems are in effect synoptic-scale eddies, which act to stir up warm and cold air masses, hence playing a large part in the transport of heat from the tropics towards the poles. As the eddies decay, the eddy kinetic energy is partly converted into zonal kinetic energy, especially at lower levels as the eddies 'barotropise' the flow. The storm track cyclones and anticyclones are therefore the same transient eddies which give rise to the eddy-driven jets. We are left with a very complex coupled system: individual storms are strongly steered by the slowly-varying large-scale flow (Lau 1988; Branstator 1995), and yet they also help to reinforce it. In this way the jet stream / storm track system can be considered 'self-maintaining' (Hoskins and Valdes 1990; Robinson 2006).

In the middle panels of Figure 1 we show a summary of the Northern Hemisphere storm tracks. Near the surface, the transient eddy activity is concentrated over the ocean basins, in the Pacific and Atlantic storm tracks, where surface friction is lowest. At upper levels, however, eddy activity continues across North America so that the two storm tracks are joined. This indicates the presence of transient

† The atmosphere is a baroclinic fluid, meaning that the density depends on both temperature and pressure. This means that temperature can vary along a surface of constant pressure, and instability which can only exist because of these horizontal gradients of temperature is called baroclinic instability. In flows of the scale of weather systems or larger the wind field is, to a good approximation, in 'thermal wind balance', so that the horizontal temperature gradients are related to vertical gradients in the horizontal wind. In contrast, a barotropic flow is one in which density is a function of pressure alone, so that temperature does not vary along a pressure surface and the (balanced part of the) horizontal wind does not vary with height.

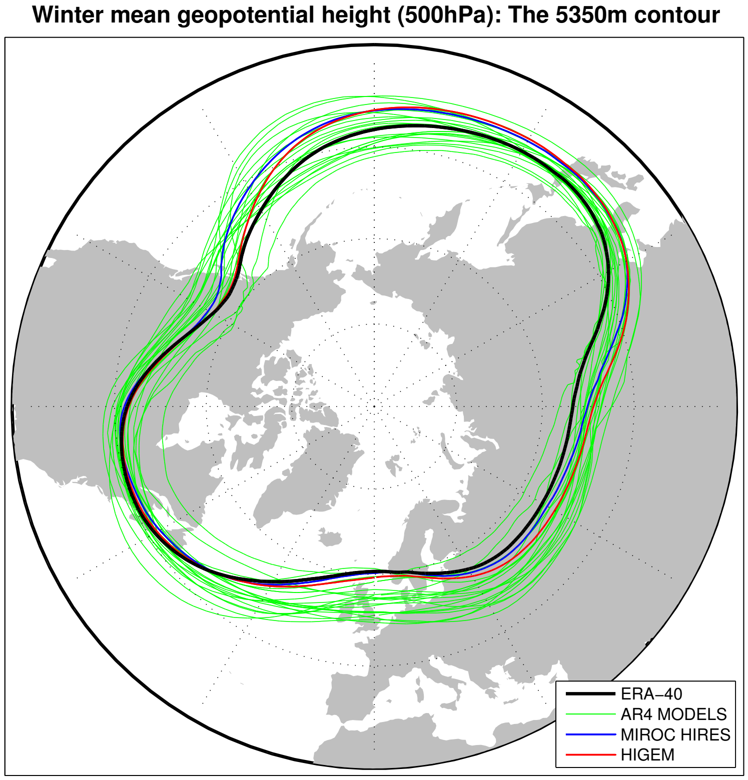

Figure 2. One contour of winter (DJF) 500 hPa geopotential height from a variety of sources. The black contour is from the ERA-40 reanalysis for winters 1957/58 - 2001/02. Each of the green contours is from one of the climate models used in the Intergovernmental Panel on Climate Change Fourth Assessment Report (IPCC AR4), using years 1960-99 of the 20th century climate runs (20C3M) from the CMIP3 archive. The models used are as in Woollings (2008). Of these, the high-resolution Miroc model is highlighted in blue. Also shown is the high-resolution HiGEM model (Shaffrey *et al.* 2009) in red, using 50 years of the present day control run.

Rossby wave packets propagating from the Pacific into the Atlantic, influencing the jet stream (*e.g.* Franzke *et al.* 2004), and ultimately dissipating their energy near Europe. In summer the storm track is weaker and slightly further poleward, although in the East Atlantic the latitude is virtually unchanged, so that the storm track points directly at the British Isles in summer as well as winter.

(*c*) *The stationary waves*

Rossby waves consist of large masses of air moving horizontally, largely northward and southward. As they do so they develop anomalous relative vorticity to balance the change in planetary vorticity, and these anomalies influence neighbouring air masses in such a way as to propagate the original disturbance towards the west, *i.e.* upstream. Rossby waves of various scales are ubiquitous in the atmosphere, and one important subset of these are the stationary waves. The zonal asymmetries of the atmospheric circulation can be interpreted as the stationary Rossby wave response to asymmetries in the surface forcing, in particular the orographic forcing from mountain ranges and the pattern of heating associated with sea surface temperatures (SSTs) and the configuration of the continents. The jet streams are not zonally symmetric, and the transient heat and momentum forcing which drives the eddy-driven jet stream also contributes to explaining the observed asymmetries. See Held *et al.* (2002) for a comprehensive review of stationary wave theory and modelling.

The stationary waves are seen clearly in contours of climatological streamfunction, as shown in the left panels of Figure 1. The (generally dominant) rotational part of the flow is directed parallel to contours of streamfunction; contours displaced towards the equator indicate a trough, or cyclonic anomaly, whereas those displaced poleward, such as those over western Europe and Alaska, indicate a ridge, or anticyclonic anomaly.

The thick black line in Figure 2 shows just one contour of geopotential height at 500 hPa, which behaves in a manner similar to streamfunction. The contour is displaced south over the western Atlantic, and then follows the eddy-driven jet stream as it tilts southwest-northeast across the Atlantic. This tilt is an important part of the stationary wave pattern, and arises due to a combination of factors, in particular the configuration of the Rocky mountains and the east coast of North America (Brayshaw *et al.* 2009). The tilt of the jet stream and the ridge over Europe are both important factors in characterising European climate, yet the AR4 climate models, shown in green, almost all underestimate these features. This tendency to simulate an overly zonal flow is an enduring problem for climate models. Such systematic biases are important reasons why European climate is particularly uncertain. However, it is encouraging that two relatively high resolution models (Miroc 3.2 in blue and HiGEM in red), match the observed flow very well. This gives hope that improving global models by increasing their resolution could reduce this problem considerably.

(*d*) *Ocean circulation*

The Meridional Overturning Circulation (MOC) transports warm water northward near the surface in the Atlantic Ocean. This results in mild SSTs at relatively high latitudes which contribute to keeping Europe significantly warmer than other regions of the same latitude†. When the warm water reaches the northernmost regions of the Atlantic it is cooled and so becomes denser and sinks, to return south at depth. Many of the key physical processes important for the MOC occur at very small scales in the ocean and so are universally poorly represented in current models (*e.g.* open ocean convection, dense overflows, boundary currents and eddies).

(*e*) *Low-frequency variability*

The extratropical atmospheric circulation exhibits a rich array of low-frequency variability, such as that associated with Rossby wave-trains. However, in the Atlantic sector the largest variability is due to north-south shifts of the eddy-driven jets. This is the variability which is described by meridional dipoles in sea level pressure, or geopotential height, such as the North Atlantic Oscillation (NAO). During the positive phase of the NAO, the eddy-driven jet is shifted to the north and intensified, with a stronger southwest-northeast tilt (*e.g.* Woollings *et al.* 2009). The zonally symmetric Northern and Southern Annular modes (NAM and SAM; Thompson and Wallace 2000) describe similar variability, although the zonal asymmetry of the climatological flow in the Northern Hemisphere complicates this interpretation (Ambaum *et al.* 2001).

The NAO, generally defined by the pattern of anti-correlated variations in sea level pressure over Iceland and the Azores, is well known to have a very strong

† The southwest-northeast tilt of the eddy-driven jet stream also contributes, by advecting warmer air from further south (Seager *et al.* 2002), although the MOC heat transport itself contributes to this tilt (Wilson *et al.* 2009).

influence on European winter climate (Wanner *et al.* 2001; Hurrell *et al.* 2003)‡. Of particular importance is the strong variability on decadal timescales. There is still much debate over the origin of the dramatic 'trend' of the NAO from strong negative conditions in the 1960s to strong positive conditions in the 1990s, even though it has since returned to a more normal state. The NAO does not in fact show a robust long term trend (Cohen *et al.* 2005; Franzke 2009), and the strongest trends in sea level pressure are located further east, reflecting a trend towards deeper penetration of the westerlies into northern Europe (van Oldenborgh *et al.* 2009). The NAO trend from the 1960s to 1990s is still of interest, however, because of the magnitude of the circulation change experienced over just a few decades. Several candidates have been suggested as influences on this trend, including tropical SSTs (Hurrell *et al.* 2004), stratospheric water vapour (Joshi *et al.* 2006) and of course anthropogenic climate change (Gillett *et al.* 2003). It remains a concern that climate models are in general unable to simulate changes as large as that observed, either as natural variability, or as the response to greenhouse gas forcing (Gillett 2005), although there are exceptions (*e.g.* Cooper and Gordon 2002; Osborn 2004; Selten *et al.* 2004; Raible *et al.* 2005). There are two significant concerns: not only do we not understand the causes of the dramatic variations which occurred over recent decades, but we are also left in doubt over the skill of current models in simulating the dynamics of the jet stream.

The rightmost panels of Figure 1 show the standard deviation of monthly mean streamfunction anomalies. This shows that variability is particularly high at the downstream ends of the Pacific and Atlantic storm tracks, where the eddy-driven jets dominate. Further upstream the subtropical jets are very strong and this inhibits the eddy-driven jets (Nakamura and Sampe 2002; Lee and Kim 2003; Eichelberger and Hartmann 2007). Yet it is the eddy-driven jet which is most variable, driven by variations in the transient eddy forcing (Thompson *et al.* 2002; Vallis and Gerber 2008). Furthermore, the diffluent flow makes the exit region of the Atlantic jet particularly conducive to transient wave-breaking and the associated

‡ Similar variability exists in summer and is also associated with strong decadal variability and impacts (Folland *et al.* 2009).

forcing (Gerber and Vallis 2009). Blocking is also common in these areas (Tyrlis and Hoskins 2008), contributing to the high standard deviations.

These maps show that in both seasons western Europe has one of the most variable climates of any land area in the Northern Hemisphere†. Europe is directly affected by variations in the jet stream, and these in turn have many different influences, such as the remote effects of El Niño-Southern Oscillation (ENSO; *e.g.* Brönnimann 2007) and the Madden-Julian Oscillation (Cassou 2008), decadal variations of the AMO (*e.g.* Folland *et al.* 2009) and, as described below, changes in the stratosphere. This myriad of influences means not only that the climate of Europe is highly variable, but it is often particularly hard to trace the origins of a given event back to a particular source.

(*f*) *Blocking*

The term 'blocking' describes a weather pattern in which the prevailing westerly winds and storms are blocked by a persistent and stationary anomaly, generally an anticyclone. Blocks persist for longer than a few days (which is the typical lifetime of storms) and have been known to persist up to several weeks (Tyrlis and Hoskins 2008). It is this persistence which leads to strong impacts, such as extended cold snaps in winter as the land surface cools under cloudless skies and often easterly winds. In summer, blocking leads to long dry spells and heatwaves. There are several different methods used to identify blocking, and the results of these indices can differ significantly. However, all methods identify Europe as a region of frequent blocking, and often as the region of most frequent blocking.

Figure 3 shows two examples of winter blocks. The first, in winter 1992/93, illustrates the dynamical mechanism underlying blocking. The normal north to south gradient in streamfunction has been reversed by a large-scale anticyclonic (clockwise) wave-breaking pattern. This is the signature of the irreversible breaking of synoptic scale Rossby waves which lead to formation of blocks (Pelly and Hoskins 2003). In this case a large mass of subtropical air has been moved up to higher latitudes over Northwest Europe, where it constitutes an anticyclonic anomaly.

† Note that the standard deviation over Europe is similar to that over the Pacific coast of North America, even though the mean flow is significantly weaker.

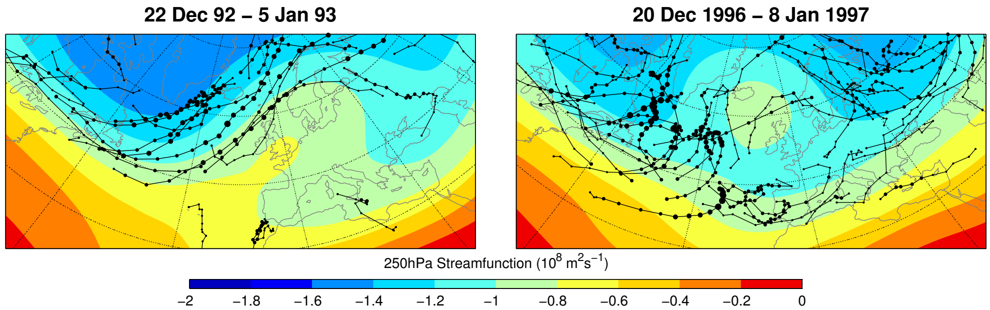

Figure 3. 250 hPa streamfunction averaged over the two periods shown. Also shown are all the cyclone tracks which overlap this period by at least one day. Cyclone tracks were derived by identifying and tracking maxima in vorticity at 850 hPa, using the method of Hoskins and Hodges (2002) applied on a regional Atlantic grid as in Woollings *et al.* (2009). Dots are marked every 6 hours along each track and are proportional to the intensity of the cyclone. All data are from the ERA-40 reanalysis.

The storms are diverted north around the block, generally following the contours of the time-mean flow. In the second example, this time for winter 1996/97, the mass of subtropical air has been cut off and isolated from the subtropics, but still persists over the Northeast Atlantic for many days. These two examples also show that the local impacts of blocking depend critically on the location of the system; in the 1992 event, much of the British Isles is still under mild southwesterly flow, whereas in the 1996 event it lies under the northeasterlies bringing cold air from Scandinavia.

(*g*) *The stratosphere*

The stratosphere is the layer of the atmosphere around 10-50km above the surface where temperature increases with height, so that an air parcel is very stable to vertical perturbations. This, along with its relatively small mass, means that for many purposes its effect on tropospheric flow can be neglected. However, there has been much interest in recent years in the possibility that variations in the stratospheric polar vortex can propagate down into the troposphere and lead to shifts in the tropospheric jet streams associated with the annular modes (Baldwin and Dunkerton 2001). There are, in fact, dramatic variations in stratospheric flow during stratospheric sudden warming events, which are periods when the stratospheric polar vortex is weakened, distorted and shifted away from the pole (*e.g.* Limpasuvan *et al.* 2004). However, it should be remembered that these are themselves forced by tropospheric variability (Polvani and Waugh 2004).

Although the influence of the stratosphere is often linked to the hemispherically symmetric NAM, the NAM itself is dominated by variations in the Atlantic (Feldstein and Franzke 2006), and changes in the stratosphere have the largest influence in the Atlantic sector (Deser 2000; Limpasuvan *et al.* 2004). Once again Europe is particularly strongly affected (Douville 2009), especially since there is evidence that the stratosphere is involved in the teleconnection of ENSO influence to Europe in late winter (Bell *et al.* 2009).

3. Future European climate

We begin this section by presenting an illustration of the spread in circulation responses in the climate models contributing to the IPCC AR4. Figure 4 shows the difference in winter mean Z500 (geopotential height at 500 hPa) between years 2060-99 of the SRESA1B scenario and 1960-99 of the 20th century control runs. All of the patterns are robust to sub-sampling of the time period, showing that natural variability does not make an important contribution to the differences shown. There is some agreement between the different models, with several predicting a decrease in heights (*i.e.* a cyclonic anomaly) over the northern North Atlantic and several predicting higher heights over the Mediterranean. However, as shown by van Ulden and van Oldenborgh (2006), there is still considerable spread, with some models predicting markedly different changes. Even among models predicting a cyclonic response over the North Atlantic there is important variation. Compare, for example, the medium and high resolution versions of the Miroc model, which place the UK under westerly and southerly anomalies respectively.

The varied circulation responses of the different climate models should not be too surprising given the findings of Sigmond *et al.* (2007), who showed that the responses of the extratropical zonal winds and stationary waves are not robust to changes of resolution or physics parameters in atmospheric models.

The spread in circulation responses is only of practical concern to the extent that it leads to uncertainty in quantities such as temperature, precipitation and sur-

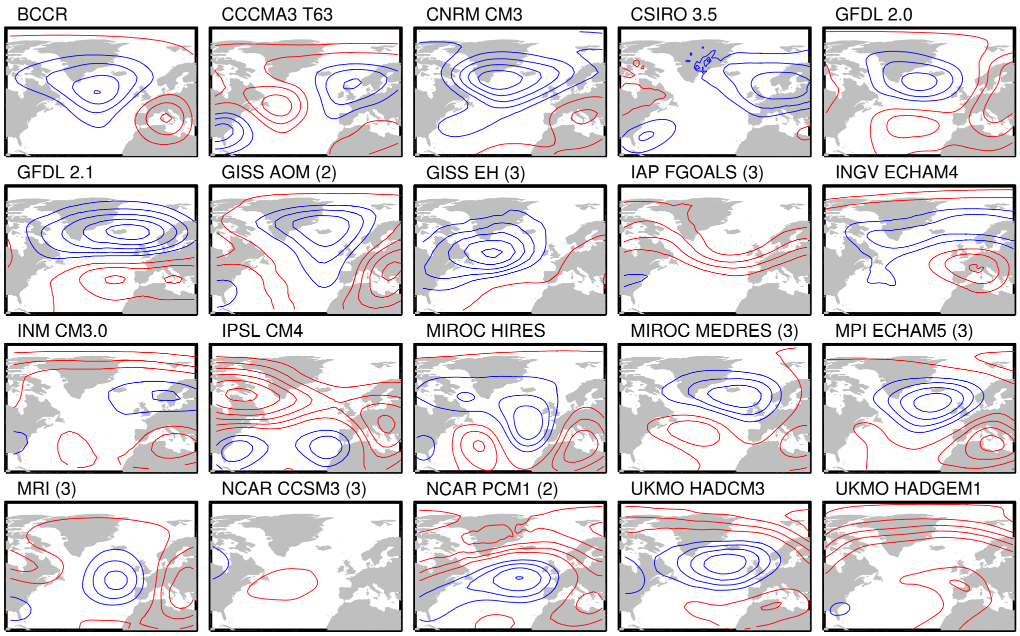

Figure 4. The difference in DJF 500 hPa geopotential height (Z500) between years 2060-99 of the SRESA1B scenario and 1960-99 of the 20th century control runs (20C3M) for 20 models from the CMIP3 archive. The differences are contoured every 10m with negative contours in blue and the zero contour omitted. Where possible, several SRESA1B ensemble members have been averaged and the number of members is indicated in brackets. See Randall *et al.* (2007) for details of the models. Note that the global mean Z500 difference, representing the average lifting of the 500 hPa surface in a warmer world, has been removed from each field for illustrative purposes.

face wind. Several studies have shown that model spread in circulation over Europe does indeed contribute to such uncertainty (*e.g.* van Ulden and van Oldenborgh 2006; Christensen *et al.* 2007; Boé *et al.* 2009). Fortunately other processes influencing European climate appear more robust, such as the change in precipitation associated with the increased moisture holding capacity of a warmer atmosphere†. Kendon *et al.* (2010) suggest that this mechanism is dominant in determining future changes in precipitation over Europe, and that changes in atmospheric circulation are generally not important enough to change the sign of the precipitation response.

† Although note that while the sign of this influence is robust there is considerable spread in the size of the associated precipitation change. This is at least partly due to the still considerable spread between models in the magnitude of the global mean surface temperature response.

However, while the circulation changes predicted by current models are relatively weak, the concern is that the potential changes may be underestimated due to systematic model biases.

In the previous section we described some of the key dynamical components of the general circulation which influence European climate. We now revisit each of these to review the evidence for how they may change in response to greenhouse gas forcing, and how they contribute to uncertainty over future European climate.

(a) *Jet streams*

In general, climate models predict that the jet streams will move poleward in response to greenhouse gas forcing (Meehl *et al.* 2007). There is evidence for an expansion of the tropics in the model simulations, suggesting that the subtropical jets will move polewards. Lorenz and DeWeaver (2007) show maps of the multi-model mean change in low-level wind, which clearly resemble a poleward shift everywhere except, interestingly, the North Atlantic in winter. Because this is the low-level wind, it is clear that the eddy-driven component of the flow is also changing.

In the Southern Hemisphere midlatitudes the change in zonal wind is equivalent barotropic, *i.e.* its sign does not change with height. This response is a clear shift of the eddy-driven jet, and as such projects very well onto the vertical structure of the SAM. However, in the Northern Hemisphere, the situation is more complicated. The response in zonal wind is baroclinic, and reflects the presence of horizontal temperature gradients, in particular associated with enhanced warming at low-levels over the North Pole (Woollings 2008). As described in Meehl *et al.* (2007), the projected warming is not uniform throughout the troposphere. It is enhanced at upper levels in the tropics due to enhanced latent heating in a warmer, moister world. It is also enhanced at low levels over the Arctic due to sea-ice and other feedbacks. The horizontal temperature gradients in the response must be accompanied by vertical wind shear according to thermal wind balance.

Potential mechanisms for the poleward shift of the jet streams generally involve the transient eddies, as discussed in the next section. For now, we simply note that there is still considerable spread between the predictions of current models for the

North Atlantic jet, as seen in Figure 4†, and in fact for the Northern Hemisphere in general (Miller *et al.* 2006). Different jet stream responses such as these are an important factor in the spread between models in European climate projections.

(*b*) *Storm tracks*

Given that the transient eddies are driven by the equator to pole temperature gradient, the pattern of atmospheric temperature change implies opposing effects on the Northern Hemisphere eddies (Held 1993). The temperature gradient is strengthened at upper levels and weakened at lower levels, though the spread between models is still very large due to uncertainty over the temperature in both the tropical and polar regions (Rind 2008). In general, the AR4 models predict that the storm tracks will shift polewards in the future, in line with the change in the jet streams (Yin 2005). This is most consistent with the upper level change, where the region of strongest baroclinicity shifts poleward.

This is, however, a classic chicken and egg problem, due to the coupling between the jets and the eddies. Does the temperature change determine the region of strongest baroclinicity and hence the location of the storm track, or do the eddies themselves change, altering the eddy-mean flow forcing and so altering the temperature structure? The answer is presumably that both of these are true, but to different extents in different models. Several studies suggest that changes in the eddies themselves are important, for example in changes to their phase speed (Chen *et al.* 2008), their vertical structure as the tropopause lifts (Lorenz and DeWeaver 2007), or their refraction as background flow gradients change (Simpson *et al.* 2009). Furthermore, changes in the atmospheric heat transports in the model simulations do indeed contribute to the steady temperature response, in particular the amplified polar warming (*e.g.* Rind 2008).

While a poleward shift of the storm tracks in future simulations seems clear in the zonal mean, especially at upper levels, there are considerable regional variations from this. In particular, the response in the North Atlantic often resembles instead a zonal extension of the storm track towards Europe (Ulbrich *et al.* 2008). There is still considerable model spread. For example, Laîné *et al.* (2009) analyse the output from

† See also Osborn (2004) and Stephenson *et al.* (2006).

two climate models in detail, showing that they predict remarkably different North Atlantic storm track changes in the future. As discussed in section 3d, differences in the MOC response may be a key factor in this.

Another factor unique to the North Atlantic is the orientation of the North American coastline, so that the land-sea contrast in predicted warming affects meridional temperature gradients in the key region at the start of the storm track where systems are growing. The sharp SST contrast across the Gulf Stream also has an influence on the genesis of the storm track (Woollings *et al.* 2009), yet this is not resolved well in current climate models. The reduction in sea-ice in a changing climate has an important effect on the storm track in some models (*e.g.* Seierstad and Bader 2008).

The role of moisture in storm track changes is also an area in need of further work. It might be expected that increased moisture in a warmer world would provide additional energy for storm development, but this does not appear to be an important factor, at least in the model simulations studied by Bengtsson *et al.* (2009). However, this could be because the processes responsible for the moist feedback are not represented well in models (Catto *et al.* 2009). It could also reflect an increase in latent heat transport in a moister atmosphere, which implies that weaker storm activity is needed to accomplish the same poleward energy transport (Held 1993). More detailed studies of the energetics of storm track changes, such as that of Boer (1995), would be very useful.

There is still much disagreement over the issue of future changes in storm intensity. Some studies, such as Lambert and Fyfe (2006), suggest there will be a reduction in total storm numbers but a higher occurrence of intense storms. However, other studies see no such change (*e.g.* Bengtsson *et al.* 2009; Sterl *et al.* 2009). It appears that differences in storm identification methods are still large factors in this, for example whether changes in the large-scale background flow are correctly accounted for. Different methodologies can give quite different impressions of the response, even in the same model data (*e.g.* Pinto *et al.* 2007). Ulbrich *et al.* (2009) give a thorough review of such issues.

Yet further uncertainty arises due to the existence of a rich structure of very small scale features within storms, such as sting jets for example (Browning 2004).

High impact mesoscale features such as these are particularly important in mature systems, as found near the end of the storm track over Europe, yet are not resolved even in relatively high resolution regional climate models.

(*c*) *Stationary waves*

There have been surprisingly few studies attempting to interpret climate model projections from the perspective of the stationary waves. A recent example is the study of Brandefelt and Körnich (2008), who analyse stationary waves in the AR4 models. Although there are encouraging signs that the models are coming into agreement in some regards, there are considerable differences which result in large uncertainties for regional climates. There is also disagreement between different studies over the mechanisms responsible for stationary wave changes, although this may partly reflect the different behaviours of different models. For example, Joseph *et al.* (2004) found a general reduction in stationary wave amplitude which could be primarily attributed to changes in the zonal mean state. However, in an earlier version of the same model, Stephenson and Held (1993) had found this to play only a minor role, with the effects of latent heating and the transients being dominant. Brandefelt and Körnich (2008) found that, on average, the changing zonal mean state contributes on the order of 50% of the total stationary wave change.

(*d*) *Ocean circulation*

There is much concern over the possibility that climate change could affect the formation of dense water at high latitudes, and so significantly weaken the MOC in just a few decades (Meehl *et al.* 2007), cooling Europe by up to 3 °C (*e.g.* Vellinga and Wood 2002). However, climate models predict that the MOC will weaken gradually, but not collapse, in response to anthropogenic forcing, mostly because of the increase in temperature and hence decrease in density at high latitudes (Gregory *et al.* 2005). The weakening of the MOC leads to reduced heat transport into the North Atlantic and a significant reduction of the anthropogenic SST increase. This ameliorates the warming projected for western Europe, but does not reverse it (Meehl *et al.* 2007). However, there is a considerable spread in the size of the MOC reduction predicted by the different models. This presumably contributes to the spread in temperature projections for Europe, but there has been little or no attempt to quantify this contribution. In addition, the effects of meltwater fluxes from the Greenland ice sheet are not included in most models, which adds to the uncertainty. Regional sea-level changes are also an important predicted consequence of MOC weakening, which would mostly affect Europe and the east coast of North America (Kuhlbrodt *et al.* 2009)†.

The weakening of the MOC leads to a minimum in warming in the North Atlantic, at a similar latitude to the northen UK (Meehl *et al.* 2007). This changes the surface meridional temperature gradients which are crucial for storm development, increasing the gradient to the south of the minimum in warming and decreasing it to the north. As described in section 3b, many models predict an extension / intensification of the North Atlantic storm track, rather than a poleward shift. This seems to be consistent with the increased baroclinicity associated with the SST change (Bengtsson *et al.* 2006; Laîné *et al.* 2009), and is similar to the storm track response in an idealised simulation of a collapse of the MOC (Brayshaw *et al.* 2009). It seems likely that the varied MOC response of models also contributes to the spread in the response of the North Atlantic storm track, but again this effect has yet to be quantified.

The ocean general circulation components of climate models do not have sufficient resolution to represent many ocean currents well. van Oldenborgh *et al.* (2009) describe some of the systematic biases resulting from this in the Atlantic basin, and suggest that these biases result in the models underestimating recent Atlantic warming trends. This work also suggests that any effects of modelled MOC changes will likely be located in the wrong areas of the North Atlantic.

(*e*) *Low-frequency variability*

Paradoxically, many studies examining changes to ‘modes’ of variability essentially describe changes in the mean state of the flow. This is especially true of studies of the mean NAO/annular mode response, as these responses are descriptions of the change to the eddy-driven jet streams. Palmer (1999) suggested the hypothesis that the mean response of the climate system to forcing should resem-

† See Katsman *et al.* (2008) for a thorough discussion of future European sea level rise.

ble the dominant patterns of natural variability, which should themselves remain unchanged. However, there is evidence that some patterns of variability do change in climate change simulations (*e.g.* Ulbrich and Cristoph 1999; Brandefelt 2006). As suggested by Branstator and Selten (2009) a two-way interaction between the variability and the response seems likely: the dominant patterns of variability do strongly shape the mean response, but they are themselves altered because of the mean state changes. For example, if the dominant jet variability is north-south shifting, and the response to forcing is for the jet to move north, then the spatial pattern of variability will shift north even though the response to forcing is clearly of the same nature as the unforced variability.

There are indications from models that the nature of low-frequency variability will change in the future. For example, Coppola *et al.* (2005) and Woollings *et al.* (2009) showed examples of how the distribution of NAO variability changes in climate models, in both cases with the positive phase of the NAO becoming more dominant. There is potential for much more work in this area in the future.

Much of the analysis of variability in climate model simulations has focused on that part of the variability which is associated with a fixed spatial pattern, such as the NAO. There is perhaps a need for a more general look at variability. Important examples are some of the recent 'extreme seasons' experienced in western Europe in particular. Several of these, such as the summers of 2003 (Black *et al.* 2004) and 2007 (Blackburn *et al.* 2008), were characterised by large-scale stationary Rossby wave-trains of incredible persistence. Persistent events such as these have led to devastating impacts, as several successive storms are steered towards the same areas, while other regions experience drought and heatwaves. However, there is no reason to expect the same pattern to occur on different occasions, unless some geographically fixed feature acts to lock the phase of the wave. Whether the occurrence of persistent heat waves, droughts or floods will change in the future is of huge importance, yet the large-scale dynamics responsible for such events has received relatively little attention. Unfortunately, there also remain significant concerns over the ability of current models to adequately represent observed variability (*e.g.* Lucarini *et al.* 2007). van Oldenborgh (2007) studied the unusually warm European autumn of 2006, concluding that unless it was simply a very rare event, climate models are underestimating circulation changes such as those which helped to cause this event.

(*f*) *Blocking*

There have been few studies analysing the predictions of climate models for changes in blocking. This may be because of a general perception that the representation of blocking in climate models is poor in any case. In fact, blocking is often cited as a key reason why current climate projections for Europe may ultimately prove to be wrong.

There is indeed a historical tendency for both weather (Tibaldi and Molteni 1990) and climate (D'Andrea *et al.* 1998) models to seriously underestimate the occurrence of blocking. However, the most skillful of current climate models perform much better. In fact, as suggested by Scaife *et al.* (2009), the diagnosed bias in blocking in these models largely reflects errors in the climatological mean state rather than the occurrence of blocking episodes themselves. Essentially, biases in the large-scale mean state project onto the indices used to identify blocking.

In Figure 5 we apply a similar analysis to that of Scaife *et al.* (2009) to the change in blocking predicted by the climate model ECHAM5. Compared to the 20th century control run, ECHAM5 predicts a reduction in blocking occurrence in the late 21st century, especially over the western sides of the two ocean basins. But how much of this is actually a change in blocking behaviour, and how much is just a reflection of the change in the mean state? To estimate this, the blocking statistics for the 20th century run were re-calculated after removing the time-mean of the data and replacing it with the time mean from the 21st century ensemble. At most longitudes, the resulting blocking occurrence is very similar to that in the 21st century ensemble, suggesting that the index is primarily measuring changes in the mean flow rather than changes in blocking.

These results are less surprising given that around 70% of the days we refer to as blocked are not statistically unusual compared to a simple red noise model (Masato *et al.* 2009). The mean value around which the red noise model varies is of crucial importance for the occurrence of 'blocking' as diagnosed by the blocking index. However, it is the particularly long lasting events (*e.g.* events longer than

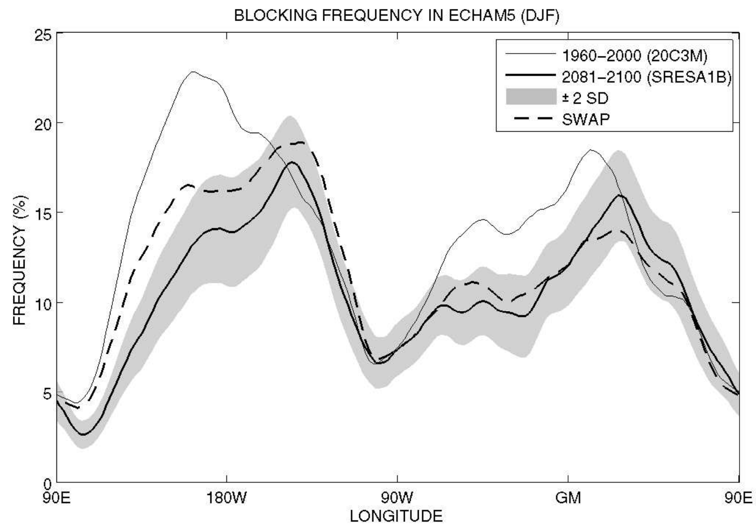

Figure 5. Frequency of occurrence of wintertime (DJF) blocking in the ECHAM5/MPI-OM coupled climate model, using simulations from the CMIP3 archive for the 20th century (20C3M) and 21st century (SRESA1B; using two ensemble members). The dashed line uses a hybrid dataset formed by taking the 20C3M data, removing the time mean and replacing it with the time mean of the SRESA1B data. The shading indicates the ± 2 standard deviation range of the SRESA1B blocking frequency, estimated by bootstrapping with 10000 trials. Blocking was diagnosed using the instantaneous D'Andrea *et al.* (1998) version of the Tibaldi and Molteni (1990) index, with $\phi_0 = 60\,^\circ$N, $\phi_n = 78\,^\circ$N, $\phi_s = 42\,^\circ$N and $\Delta = -4, 0, 4\,^\circ$. Geopotential heights were estimated from zonal wind assuming geostrophy. The frequencies have been smoothed with a 5 point running mean before plotting.

at least 10 days) which are most unusual compared to the red noise model. This suggests that attention should be focused on these since they are, after all, more clearly associated with persistent dynamical events and are also likely to give rise to the strongest impacts. Such events are, of course, rarer than their shorter-lived relations but, as extreme events, their contribution to defining the character of regional climate may well be greater than would be expected given their occurrence.

These arguments suggest that the representation of blocking may not be such a major concern for the prediction of future climate. In the example of Figure 5 there is not necessarily much change in blocking events themselves in response to the forcing. However, using the same model Sillmann and Croci-Maspoli (2009) show that blocking becomes longer lasting and more intense in future model simulations, even though the frequency of blocking decreases. Such a nonlinear change suggests that in this case there are changes in the nature of blocking itself, which would have important consequences.

(*g*) *The stratosphere*

Ever since the work of Shindell *et al.* (1999) there has been considerable interest in the possibility that stratospheric dynamics may play an important role in the response of the tropospheric jet streams to anthropogenic forcing. This seems particularly clear in the Southern Hemisphere where changes in stratospheric ozone are also important (Gillett and Thompson 2003). Despite this, many of the models contributing to AR4 still had a relatively poor representation of the stratosphere (Cordero and Forster 2006).

Other model experiments have confirmed that there can be significant dynamical changes in the Northern Hemisphere stratosphere in response to anthropogenic forcing. For example, models generally predict that the Brewer-Dobson circulation will increase in strength (Butchart *et al.* 2006) which will affect temperatures in the lower stratosphere and so also the winds near the tropopause. Some models predict an increase in the occurrence of sudden stratospheric warmings (*e.g.* Charlton-Perez *et al.* 2008), and in some cases the effect of this is strong enough to alter the tropospheric response to greenhouse gases (Huebener *et al.* 2007; Bell 2009). The relatively small number of 'stratosphere-resolving' models means there is still large uncertainty over these effects. In addition, even these models underestimate the occurrence of sudden warmings (Charlton *et al.* 2007), so the importance of these changes may still be underestimated. On the other hand, Sigmond *et al.* (2008) gave an example of a stratosphere-resolving model in which the altered tropospheric response could be shown to arise due to changes in a parameterisation scheme rather than in the vertical resolution, suggesting that the results of such models may not always be as clear as they appear.

4. Concluding remarks

To conclude, the climate of Europe is unique in several regards, and as such it poses a unique challenge for climate prediction. Europe is affected particularly strongly by many different influences, from the ocean circulation to variations in the stratosphere. In both winter and summer it has one of the most variable climates on Earth. To some extent it is the unique configuration of the Atlantic jet stream and storm track, with a strong southwest-northeast tilt and a clear separation between the two jets, which is responsible for these traits. Europe is arguably the only place in the world where a storm track and a jet stream end, and it is one of the most important regions of the world for blocking.

In many regards the climate changes currently predicted for Europe are more benign than those elsewhere in the world, partly due to the cooling effect of a weakening MOC in models. However, there are already indications that model predictions underestimate the current observed warming, due to a combination of circulation effects and model inadequacies (van Oldenborgh *et al.* 2009).

The spread between the projections of different models is particularly large over Europe, leading to a low signal to noise ratio. This is the first of two general reasons why European climate change must be considered especially uncertain. The other is the long list of physical processes which are very important for defining European climate in particular, but which are represented poorly in most, if not all, current climate models. In this paper we have discussed examples such as the tilt of the Atlantic jet stream, the dynamics of the stratosphere, the mesoscale features within mature mid-latitude cyclones and the small-scale processes at the heart of the ocean circulation.

Revisiting the UKCP09 projections serves to illustrate a key difference between these two reasons for uncertainty. The UKCP09 consortium have used a very sophisticated methodology to account for the varied projections of different models in calculating their uncertainty ranges. In contrast, it is not clear how to quantify the consequences of the systematic biases, *i.e.* the poor representation of certain physical processes in most, if not all climate models.

There is encouraging evidence that improving climate models, for example by increasing resolution, will lead to improvements in the ability of models to simulate

European climate. However, the recent work of Matsueda *et al.* (2009) raises the possibility that even with vastly improved resolution the climate of a model may not converge. Although this is a preliminary study with an atmosphere-only model, it does suggest that even with high resolution models in the future, there will still be significant differences between the predictions of different models. It will be important to continue to develop a range of models of different complexities and models based on innovative numerical methods. Understanding the differences between different climate models will remain as important as it is today.

I would like to thank NERC for the support of a postdoctoral fellowship and Brian Hoskins, Mike Blackburn, Rowan Sutton, Len Shaffrey, Helen Johnson and Chris Bell for many useful discussions. I am grateful to two anonymous reviewers for helpful and constructive comments. I acknowledge the European Centre for Medium-range Weather Forecasting for providing the ERA-40 reanalysis data, and the modelling groups, the Program for Climate Model Diagnosis and Intercomparison (PCMDI) and the WCRP's Working Group on Coupled Modelling (WGCM) for their roles in making available the WCRP CMIP3 multi-model dataset. Support of this dataset is provided by the Office of Science, U.S. Department of Energy.

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