Some physical drivers of changes in the winter storm tracks over the North Atlantic and Mediterranean during the Holocene

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The winter climate of Europe and the Mediterranean is dominated by the weather systems of the mid-latitude storm tracks. The behaviour of the storm tracks is highly variable, particularly in the eastern North Atlantic, and has a profound impact on the hydroclimate of the Mediterranean region. A deeper understanding of the storm tracks and the factors that drive them is therefore crucial for interpreting past changes in Mediterranean climate and the civilizations it has supported over the last 12,000 years (broadly the Holocene period). This paper presents a discussion of how changes in climate forcing (e.g. orbital variations, greenhouse gases, ice sheet cover) may have impacted on the ‘basic ingredients’ controlling the mid-latitude storm tracks over the North Atlantic and the Mediterranean on intermillennial time scales. Idealized simulations using the HadAM3 atmospheric general circulation model (GCM) are used to explore the basic processes, while a series of timeslice simulations from a similar atmospheric GCM coupled to a thermodynamic slab ocean (HadSM3) are examined to identify the impact these drivers have on the storm track during the Holocene. The results suggest that the North Atlantic storm track has moved northward and strengthened with time since the Early to Mid-Holocene. In contrast, the Mediterranean storm track may have weakened over the same period. It is, however, emphasized that much remains still to be understood about the evolution of the North Atlantic and Mediterranean storm tracks during the Holocene period.

Keywords: storm track; Holocene; climate; Mediterranean; North Atlantic

1. Introduction

(a) Motivation

Synoptic-scale weather systems (storms) dominate the winter climate of Europe and the Mediterranean. The aggregate path of these storms is known as the storm track and storm activity is strongest in the winter season. The modern-day

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storm tracks are shown in figure 1, with the North Atlantic storm track running southwest–northeast across the Atlantic, and a smaller storm track running west–east across the Mediterranean basin (see appendix A for details of the methods used to diagnose the storm tracks). A third major storm track occurs in the North Pacific basin, running almost zonally west–east (not shown).

The behaviour of the storm tracks is highly variable, particularly in the eastern North Atlantic (i.e. in the area ‘linking’ the Mediterranean and North Atlantic storm tracks). In the modern day, atmospheric variability in this region has a large impact on the hydroclimate of the Mediterranean area (e.g. Corte-Real et al. 1995; Trigo et al. 2000; Krichak et al. 2002; Dunkeloh & Jacobeit 2003; Xoplaki et al. 2004; Krichak & Alpert 2005), where the weather systems associated with the storm track provide much of the annual rainfall (the Mediterranean is characterized by warm dry summers and cold wet winters; see also figure 1a). Many areas of the Mediterranean experience high levels of water stress in the present day, and this stress is anticipated to increase in the future (Meehl et al. 2007; WGBU 2007; Mariotti et al. 2008), demonstrating the need to establish a good physical understanding of the processes controlling the evolution of these storm tracks.

Archaeological and palaeo-proxy studies suggest that changes in physical climate (particularly water stress) have been a major driver in the development and decline of ancient civilizations (the Maya of Central America (Haug et al. 2003) and the Akkadians of Mesopotamia (Weiss et al. 1993) for examples of well documented specific events globally). A wider summary of the impact of Holocene climate change in the Mediterranean can be found in Roberts et al. (submitted) and Issar & Brown (1998). There is therefore enormous cultural value in developing a deeper knowledge of regional palaeoclimates over the Holocene (defined here as the last 12 000 years).

As well as its archaeological relevance, palaeoclimate is frequently cited as a way to ‘test’ climate models against each other and against palaeo-proxy observations, e.g. the PMIP projects: see Braconnot et al. (2007a), Brewer et al. (2007), Bonfils et al. (2004) and Jolly et al. (1998). Indeed, this approach has been pursued both globally and regionally using the model data analysed in this paper (see appendix B for a discussion of the TIMESLICE experiments, and see Brayshaw et al. (in press a,b) and Black et al. (in press) for details of this comparison). In practice, however, detailed regional-scale comparison between proxy records and model data tends to be extremely difficult for many reasons, including

— model uncertainties (no model produces a perfect representation of climate, cf. the wide range of responses in PMIP-2 Mid-Holocene models over the North Atlantic; Gladstone et al. 2005);
— spatial resolution (climate models are limited in resolution by the computational power available to typically 1–3°, but palaeorecords and archaeological sites may be strongly influenced by local topography);
— temporal variability (the climate system is chaotic and contains significant low-frequency unforced or ‘natural’ variability); and
— conflicting palaeo-proxy interpretations (e.g. a change in isotopic oxygen ratios may result from changes in temperature, precipitation or moisture source).
There is therefore a need to better understand the behaviour of the Mediterranean storm track and its links to the atmospheric circulation over the North Atlantic, both for predicting future climate and for understanding the past. Although this present paper focuses on the climatological forcing of the atmosphere over the last 12,000 years, many of the dynamical arguments presented are therefore relevant to modern day climate. To this end, a combination of dynamical and process-based theory is used alongside climate model integrations to investigate how the North Atlantic and Mediterranean storm tracks may have changed during the Holocene. These are then contrasted with some of the major characteristic patterns of palaeo-reconstructions and other climate model simulations. The reader is, however, referred to Brayshaw et al. (in press a, b) and Black et al. (in press) to find a more thorough review of model versus proxy data comparison in the Mediterranean.

The paper begins with a review of the major changes in the climate forcing parameters (orbital variations, greenhouse gas concentrations, etc.) during the Holocene (§1b), followed by a discussion identifying the major meteorological drivers controlling the North Atlantic and Mediterranean storm tracks (§1c). The changes in each of the meteorological drivers is then discussed, first for the North Atlantic storm track (§2), followed by the Mediterranean storm track (§3). At the end of the discussion for each storm track, there is a summary outlining the aspects of storm track change we can be most confident in. The paper concludes with a summary discussion (§4).

To examine the changes in the storm track, two distinct sets of atmospheric climate model (GCM) integrations are used: ‘IDEALISED’ experiments and ‘TIMESLICE’ experiments. The IDEALISED experiments examine the storm

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Table 1. The TIMESLICE experiments. The ‘short ID’ column indicates the shortened experiment identifiers used in figure 4. The ‘GHG’ column indicates greenhouse gas concentrations (see figure 2 for CO₂ radiative forcings), with ‘PD’ for present day (approx. 1970) and ‘PI’ for preindustrial. ‘MH’ and ‘EH’ denote the MID-HOLOCENE and EARLY-HOLOCENE experiment groups respectively (where an average is taken across the component experiments). The ‘error’ column indicates which experiments are affected by the subtle configuration error described in the main text and appendix B.

<table>
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It should, however, be noted that during the analysis of the TIMESLICE experiments, a subtle error was discovered that affects some of the earlier time periods only (8 ka BP and earlier). This makes it difficult to interpret the TIMESLICE experiments as a simple time series of points across the Holocene. To minimize the impact of this error, two representative time-step comparisons are used:

— ‘MID-HOLOCENE’ (broadly 6–8 ka BP) versus ‘PREINDUSTRIAL’; and
— ‘EARLY-HOLOCENE’ (broadly 10–12 ka BP) versus ‘8 ka BP’.¹

The former of these comparisons is not affected by the model configuration error but does not include the effects of changes in land–ice sheet cover whereas, in the latter comparison, land–ice sheet changes are included but the experiments are all affected by the model configuration error. It is not, however, expected that the error significantly affects the discussion presented in this paper. Further discussion is provided in appendix B and a full description can be found in Brayshaw et al. (in press a).

¹Note that there are two different experiments corresponding to conditions at 8 ka BP (see table 1). MID-HOLOCENE is constructed from experiments 6 ka BP and 8 ka BP-NOICE, whereas experiment 8 ka BP (rather than experiment 8 ka BP-NOICE) is used in the comparison with EARLY-HOLOCENE.
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Consistent with the aim of this paper (to examine the atmospheric response to climate forcing during the Holocene), three main changes in climate forcing are identified on inter-millennial scales during the Holocene:

— greenhouse gas concentrations (concentrations from Jansen et al. (2007), converted to radiative forcings following Myhre et al. (1998));
— ice sheet extent over land areas (based on ICE-4G; Peltier 1994); and
— insolation (from orbital variations; Berger 1978).

These forcings are illustrated in figure 2 and discussed below.

Since the beginning of the Holocene, greenhouse gas concentrations (particularly CO₂ and CH₄) have increased, consistent with a radiative forcing of approximately 0.5–1 W m⁻² between the Early Holocene (10–12 ka BP) and

**Figure 2. Climate forcing during the Holocene.**

(a) CO₂ radiative forcing (concentrations from Jansen et al. (2007) converted to radiative forcing using Myhre et al. (1998)). (b) Change in land ice sheet cover between 12 ka BP and the preindustrial period (based on ICE-4G (Peltier 1994), units m). (c) The annual cycle in TOA insolation under present day (or preindustrial) conditions. (d) The difference in TOA insolation between 8 ka BP and preindustrial conditions. (e) Annual mean TOA insolation changes during the Holocene. (f) November to April mean TOA insolation changes during the Holocene. Units for (c–f) are W m⁻².

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preindustrial conditions (figure 2a). Comparison with expectations of modern-day anthropogenic climate change suggests that this is consistent with an increase in global mean surface air temperature (SAT) of around 0.5–1°C between these two periods (Trenberth et al. 2007).

During the Early to Mid-Holocene (6–12 ka BP), the large ice sheets developed during the last Ice Age gradually receded over northern Eurasia and northeastern America, linked to a sea-level rise of approximately 60 m (Jansen et al. 2007). A reconstruction of the difference between the ice sheets at 12 ka BP and preindustrial levels is shown in figure 2b. By approximately 6 ka BP most of this additional land–ice sheet had melted.

Perhaps the largest and most complex difference between the climate forcing in the Early Holocene and that under preindustrial conditions is associated with variations in the Earth’s orbit around the Sun. Figure 2c, d shows the differences in the seasonal cycle of insolation at the top of the atmosphere (TOA) between 8 ka BP and preindustrial conditions. Over the Northern Hemisphere, the differences tend to exaggerate the seasonal cycle of radiation at 8 ka BP whereas, in the Southern Hemisphere, there is a generally weaker seasonal cycle (these changes are up to 15–20% of the total insolation). It should, however, also be noted that the insolation differences are not completely in phase with the preindustrial seasonal cycle, e.g. peak insolation in the Northern Hemisphere occurs in June under preindustrial conditions (figure 2c) whereas the peak difference between 8 ka BP and preindustrial occurs in July (figure 2d). Furthermore, it should be noted that this phase relationship changes gradually during the Holocene. In general, however, the insolation difference acts to modulate the intensity of the seasonal cycle, with the largest difference from preindustrial conditions near 9–10 ka BP (see figure 2f).

The variations in the Earth’s orbit also act to modify the latitudinal distribution of insolation in the annual mean (figure 2e). In the Early Holocene, the tropics received 1–2 W m⁻² less insolation than under preindustrial conditions whereas the high latitudes received up to 6 W m⁻² more. When averaged over a full hemisphere, however, the total annual-mean insolation remains approximately constant during the Holocene period.

Throughout this paper, the focus is on the atmospheric response to these changes in climate forcing. To simplify this discussion, it is assumed that heat transports by the ocean circulation are constant on inter-millennial time scales during the Holocene. The ocean is therefore assumed to respond as a passive thermodynamic slab gaining or losing heat in response to changes in the surface energy budget, but without its circulation being affected by changes in the surface energy budget or mechanical forcing. Evidence from previous palaeoclimate modelling programmes suggests that the dynamical ocean response to climate forcing acts to amplify some of the changes in climate during the Holocene period, and climate models which include a dynamical ocean tend to be in better agreement with palaeo-observations (e.g. the northward extension of the West African monsoon during boreal summer in the Mid-Holocene; Braconnot et al. 2007a), which suggests that the changes in Holocene climate will tend to be underestimated in the TIMESLICE modelling framework used here (appendix B). Dynamical coupling between the atmosphere and ocean also tends to increase variability in the climate system on multi-annual time scales (in this paper, however, the focus is on changes in the mean climate rather than its
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variability). The simplistic treatment of the ocean circulation also implies that the experiments and discussion presented here adopt a ‘quasi-equilibrium’ (i.e. ‘time-slice’) perspective, rather than seeking to examine the full transient climate system response. In particular, the circulation of the ocean may adjust to changes in climate forcing over many centuries or millennia, with potentially significant impacts upon the atmospheric circulation (e.g. Lorenz & Lohmann 2004).

Despite these difficulties, to understand the full climate response it is perhaps a necessary (although not sufficient) condition that the response of the individual components is well understood. Comparison between the TIMESLICE experiments used here and the fully coupled atmosphere–ocean models of PMIP-2 suggests that useful insight into this can be gained using the simplified ocean framework described above (Brayshaw et al. in press b). Furthermore, using a thermodynamic slab ocean model is computationally much cheaper than fully coupled climate models, enabling more experiments to be performed (covering a wider range of time periods during the Holocene) while incorporating the fine spatial resolution required to study storm track processes in the Mediterranean.

(c) The basic ingredients of storm tracks

Mid-latitude storms form in regions that are baroclinically unstable (i.e. with strong horizontal temperature gradients or, equivalently, a strong vertical shear in the horizontal wind). Following Hoskins & Valdes (1990), a commonly used measure of baroclinicity is the maximum Eady growth rate:

\[
\sigma = -\frac{0.31g}{N} \frac{\partial \Theta}{\partial y} = \frac{0.31f}{N} \frac{\partial \bar{u}}{\partial z},
\]

where \( f \) is the Coriolis parameter, \( N \) the static stability, \( \Theta \) the potential temperature, \( \Theta_0 \) a reference potential temperature profile and \( u \) the zonal wind. Overbars are used to indicate a time-mean.

In general, the baroclinic region lies on the poleward flank of the subtropical jet (around 30–40° latitude), associated with the limit of the poleward energy transport by the atmospheric Hadley cells. The storms themselves are a fundamental part of the climate system’s poleward energy transport mechanism, with storm activity tending to reduce the baroclinicity upon which they formed (Simmons & Hoskins 1978).

The poleward transport of heat is ultimately driven by latitudinal differences in insolation (figure 2c shows that the equator receives more insolation than the poles). The equator-to-pole gradient in insolation is smaller in the Early to Mid-Holocene than it is under preindustrial conditions (figure 2e) and, as a consequence, one might expect that the poleward transport of energy in the climate system would have been lower in the Early to Mid-Holocene. In the mid-latitudes, storm track processes represent the dominant form of poleward energy transport (e.g. Trenberth & Caron 2001; Trenberth & Stepeniak 2004; Fasullo & Trenberth 2008) therefore suggesting that it is probable that the storm tracks of the Early to Mid-Holocene would have been somewhat weaker than those experienced under preindustrial conditions (Boer 1995).

Storm development is strongly affected by atmospheric moisture content, with latent heating associated with condensation acting to both intensify storms (e.g. Wernli et al. 2002), and maintain the geographical structure of the storm
tracks (Hoskins & Valdes 1990). During boreal winter in the Early to Mid-Holocene, the atmospheric moisture content can be expected to be lower than that under preindustrial conditions (consistent with generally lower near surface temperatures associated with weaker insolation; figure 1f and see also Brayshaw et al. in press a, b). The reduced atmospheric moisture content may be particularly pronounced over land areas (as they cool more rapidly during winter than the oceans owing to differences in their surface heat capacities), which may also act to focus storm activity and development more strongly over the ocean basins (e.g. Inatsu et al. 2002; Cash et al. 2005; Brayshaw et al. 2009). In contrast, over high-latitude oceans the annual-mean insolation in the Early to Mid-Holocene is greater than that under preindustrial conditions (figure 2e) which may be associated with greater atmospheric moisture content during winter over warmer oceans with less sea ice in the Early to Mid-Holocene. The impact of moisture changes on the storm tracks is therefore a complex and challenging issue.

The structure of the Northern Hemisphere atmospheric circulation is strongly affected by zonal asymmetries in the planetary boundary conditions (e.g. Nigam et al. 1988; Valdes & Hoskins 1989; Broccoli & Manabe 1992; Held et al. 2002; Inatsu et al. 2002; Cash et al. 2005; Brayshaw et al. 2009). Such features may either act to damp storm activity (for example, land surfaces are rougher and drier than ocean surfaces; e.g. Chang & Orlanski 1993; Brayshaw et al. 2009) or locally enhance the rate at which the baroclinicity of the atmosphere is restored (e.g. locally strong surface temperature gradients; Nakamura et al. 2004). Brayshaw et al. (2009, in preparation) used an atmospheric GCM to show that much of the basic structure of the North Atlantic storm track, particularly its southwest to northeast tilt across the North Atlantic Ocean, was produced by a combination of: (i) the large-scale atmospheric circulation patterns generated by the Rocky mountain range; (ii) the shape of the North American continent; and (iii) sea surface temperatures (SSTs) in the North Atlantic, with tropical land–sea distributions perhaps playing a secondary role.

Given the climatological forcings associated with the last 12000 years (discussed in §1b), and the planetary features that control the storm track’s location, it is postulated that changes in the following three properties may have had a large impact on the character of the North Atlantic storm track during the Holocene (see figure 3a):

— the tropical circulation and its impact upon the subtropical jet position and intensity;
— the surface contrast at the east coast of North America; and
— SSTs and sea ice changes in the North Atlantic.

The mechanisms associated with each process and their impact upon the storm track are discussed in §2.

The Mediterranean storm track is more subtle and generally less well understood than the North Atlantic storm track. Mediterranean storms are frequently generated by the interaction of the flow associated with synoptic-scale weather systems over Europe and the complex topography in the Mediterranean basin. Examples of this process include Alpine lee cyclogenesis (Buzzi & Tibaldi 1978) and Aegean cyclogenesis (Flocas & Karacostas 1996), with a single cyclone tracking eastward over southern and central Europe being capable of...
triggering multiple cyclogenesis events over the Mediterranean (Trigo et al. 2002). Studies have also emphasized the importance of disturbances in the upper tropospheric flow for Mediterranean cyclogenesis (Flocas & Karacostas 1996; Thorncroft & Flocas 1997), which suggests that the proximity of the subtropical jet entrance over North Africa may play some role, particularly in Saharan cyclogenesis. On a ‘climatological’ level, the quasi-zonal coastline of the north Mediterranean creates a strong surface temperature front between the cold winter land mass and the warmer ocean to the south, producing a strong source of near-surface baroclinicity, which could also be expected to promote storm activity (Trigo et al. 2002).

There are therefore several processes that can be expected to influence the Mediterranean storm track in the context of Holocene palaeoclimates. These include (see figure 3b)
— the strength and position of the North Atlantic storm track;
— the surface contrast between the cold land mass and the warm waters at the northern coastline of the Mediterranean; and
— the proximity and strength of the subtropical jet entrance over Africa.

These are discussed in §3.

2. The North Atlantic storm track

The mechanisms influencing the North Atlantic storm track are discussed in four separate sections. The first two sections examine the time-mean winter tropical circulation in terms of its zonal-mean response (§2\textit{a}) and its latitudinally dependent response (§2\textit{b}). This is followed by a discussion of the influence of land–sea contrast at the coastline of North America (§2\textit{c}) and SSTs in the North Atlantic (§2\textit{d}). A summary section is provided at the end (§2\textit{e}).

(a) Zonal-mean tropical response

The tropical troposphere’s zonal-mean overturning circulation, known as the Hadley cell, is driven primarily by the differential gradient of net insolation between the deep tropics and the subtropics (Held & Hou 1980). The Hadley cell is strengthened by latent heat release associated with precipitation in the intertropical convergence zone (the ITCZ; e.g. Kim & Lee 2001) and surface drag (via Ekman pumping; e.g. Cook 2003).

A recent series of studies (Held & Soden 2006; Lu et al. 2007; Vecchi & Soden 2007) have examined the response of the tropical circulation (and the Hadley cell in particular) to the modern-day greenhouse gas forcing problem using the IPCC Fourth Assessment models. In particular, Held & Soden (2006) argue that, as near-surface atmospheric temperatures increase in response to the greenhouse forcing, so does the near-surface atmospheric moisture content (by approx. 7\% K$^{-1}$, consistent with the Clausius–Clapeyron relationship if a constant relative humidity is assumed). The precipitation produced by the same models, however, only increases at roughly 2\% K$^{-1}$. Therefore, if all the water vapour lifted by the ascent branch of the Hadley cell is returned as precipitation, then there must be a weaker vertical mass flux in the ascent branch of the Hadley cell, and an overall weakening in the tropical circulation. Lu et al. (2007) note also that there is an expansion of the width of the Hadley cell, pushing the baroclinically unstable ‘mid-latitude’ region further poleward (perhaps consistent with the simulated poleward shift in the storm tracks; Yin 2005). While there clearly remains much to be understood in this process, not least in reconciling this result with recent observations of tropical lapse rates (see Held & Soden 2006 and references therein), it is clear that most current GCM climate models exhibit this tropical circulation response to changes in greenhouse gas concentrations.

In the context of the Holocene, greenhouse gas concentrations were lower during the Early Holocene than in current climate, although the change in concentrations (from approx. 240 ppm CO$_2$ in 12 ka BP to 280 ppm under preindustrial conditions) is rather small compared with the modern-day anthropogenic signal (figure 2\textit{a} and experiment PRESDAY take ‘present-day’ levels to be 330 ppm CO$_2$
which is roughly equivalent to that in the 1970s but more recent measurements since 2000 suggest levels closer to 380–390 ppm CO2). Nevertheless, using the TIMESLICE climate integrations and following the method used by Held & Soden (2006), it is possible to see from figure 4a,b that the strength of the annual-mean Hadley circulation (as measured by the vertical mass flux) is

— weaker in the PRESDAY experiment compared with PREIND (by approx. 5%), and
— stronger in the ‘earlier’ time-slice experiments than those corresponding to periods later in the Holocene (experiment 6ka BP is approx. 2% stronger than experiment PREIND).

A similar correlation between the vertical mass flux and greenhouse gas concentration is also found if only the winter half-year is considered (November–April, not shown). Unlike Vecchi & Soden (2007), however, no clear relationship is found between the annual-mean vertical mass flux in the TIMESLICE experiments and the annual-mean area-total upward pressure velocity (i.e. the pressure velocity area-integrated over ascent regions only: compare fig. 4 of Vecchi & Soden (2007) with figure 4c in this paper). The discrepancy between these two results is not fully understood, but may indicate that changes in the spatial and seasonal distribution of insolation during the Holocene, combined with inter-hemispheric asymmetries in the surface boundary conditions, are playing a role in determining the horizontal structure of the vertical mass flux. If, however, the analysis is restricted to the winter half-year (November–April), then a weak positive relationship between the vertical mass flux and the area-total upward mass flux is found in the TIMESLICE experiments (figure 5a). Indeed, one might expect that during the Early to Mid-Holocene when there is weaker boreal winter insolation than under preindustrial conditions, that atmospheric moisture content arguments similar to that of Held & Soden (2006) would lead the mass flux of the Hadley cell to be even stronger compared with preindustrial levels.

As the storm tracks are most active in winter, it is most relevant here to consider only this half-year extended winter period. The resulting difference in the tropical Hadley cells between the MID-HOLOCENE and PREIND experiments is shown in figure 5b. The northern Hadley cell in MID-HOLOCENE can clearly be seen to be stronger and narrower than that of experiment PREIND (consistent with the discussion above), but also slightly shallower (perhaps associated with reduced latent heat release in the ascent branch owing to lower levels of atmospheric moisture content) and shifted somewhat southwards. The southward shift in the entire cell is consistent with the difference between Mid-Holocene and preindustrial seasonal insolation (figure 2f) being stronger over the northern tropics than the southern tropics (i.e. MID-HOLOCENE has stronger negative anomalies compared with PREIND in the northern tropics than the southern tropics).

The impact of these changes in the Hadley cell on the zonal-mean zonal wind in MID-HOLOCENE is shown in figure 5c. The surface easterlies in the northern tropics are strengthened and shifted equatorward in MID-HOLOCENE compared with PREIND (easterly anomalies at 5°N, 900 hPa), and there are suggestions that the Northern Hemisphere subtropical jet position is
Figure 4. Annual mean tropical circulation in the TIMESLICE simulations. (a) Estimated fractional change in vertical mass flux in the tropics ($30^\circ$N to $30^\circ$S) across the TIMESLICE simulations, where the dark grey line denotes fractional change in precipitation ($dP/P$) and the light grey line denotes the fractional change in specific humidity estimated from the change in surface air temperature assuming that specific humidity increases at $7\% K^{-1}$ ($0.07dT$; see main text for details). The fractional change in vertical mass flux ($dM/M$; solid black line) is calculated as $dM/M = dP/P - 0.07dT$ following Held & Soden (2006). Vertical lines at each time period on each of the curves indicate the standard deviation of the interannual variability. (b) Estimated fractional change in vertical mass flux ($dM/M$) compared with estimated CO$_2$ radiative forcing, the black line is a linear fit using the $\chi^2$-method. (c) Mean fractional change in pressure velocity ($dW/W$) over tropical regions ($30^\circ$N to $30^\circ$S) with upward motion compared with estimated vertical mass flux. All changes are compared with experiment PREIND.

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Figure 5. Winter half-year mean (November–April) zonal-mean tropical circulation. (a) As figure 4c, but for the winter-mean circulation. (b) Differences in the zonal mean meridional circulation in the tropics between the MID-HOLOCENE experiment and the PREIND experiment (colours, units 10^9 kg s^{-1}) and the zonal mean meridional circulation in the PREIND experiment (contours, units 10^{10} kg s^{-1}). (c) Differences in the zonal mean zonal wind between the MID-HOLOCENE experiment and the PREIND experiment (colours, m s^{-1}) and the zonal mean zonal wind in experiment PREIND (contours, m s^{-1}). Areas marked with crosses in (b, c) indicate statistical significance at the 95% level.
downward and equatorward of that in PREIND (westerly anomalies at 25°N, 400 hPa). The signal seen in the TIMESLICE experiments, combined with the theoretical arguments and modern-day studies presented above, therefore suggests that the time-mean zonal-mean Early to Mid-Holocene Northern Hemisphere winter storm tracks would be shifted southwards compared with preindustrial conditions. Evidence to support this can be seen in figure 5b,c: the eddy-driven Ferrel cell weakens in intensity and shifts southwards (figure 5b, compare its negative centre at 35°N, 500 hPa in PREIND with the positive stream function anomaly centred at 25°N, 800 hPa) and there is an equivalent barotropic southward shift in the jet between approximately 45°N and 35°N (figure 5c).

The changes are stronger in boreal winter than boreal summer, consistent with the orbitally driven differences in boreal winter insolation acting to enhance the radiative response to greenhouse gas changes whereas those in boreal summer would tend to oppose it. However, it should not be supposed that the atmospheric response to orbitally driven insolation changes matches that of the greenhouse gas concentrations, even within a limited area such as the tropics. The atmospheric response is also relatively weak when compared with the background noise of natural variability, even within the simplified ‘slab ocean’ modelling framework used for the TIMESLICE experiments (see §1b and appendix B). The weak nature of the atmospheric response signal is, however, in part because of the highly asymmetric response of the atmospheric circulation caused by asymmetries in the planetary surface. This is discussed in the following section.

(b) The latitudinally dependent tropical response

In addition to a stronger and narrower Hadley cell in the Early to Mid-Holocene, Held & Soden (2006) and Vecchi & Soden (2007) show that most climate models produce a stronger zonally asymmetrical tropical circulation when greenhouse gas concentrations are reduced. In Vecchi & Soden (2007), this response is furthermore shown to be because of internal atmospheric processes, rather than requiring dynamical ocean processes. It is therefore unsurprising that many asymmetrical aspects of the greenhouse gas response described by Vecchi & Soden (2007) can also be seen in the boreal winters of the TIMESLICE integrations described here. The full (longitudinally dependent) response to the MID-HOLOCENE climate forcing compared with that of PREIND is shown in figures 6 and 7. Some aspects of this tropical response to lower greenhouse gas concentrations include (cf. figure 6b here with fig. 7b of Vecchi & Soden (2007) and figure 6c here with fig. 8b of Vecchi & Soden 2007)

— stronger ascent over the eastern tropical Indian Ocean and the maritime continent;
— stronger descent over southern Asia, stretching westwards towards the Eastern Mediterranean;

It should, however, be recalled that Vecchi & Soden (2007) examine the annual-mean response to greenhouse gas forcings in the present day, rather than the winter ‘half year’ mean response to the combined Holocene climate forcing in the TIMESLICE experiments.

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Figure 6. Vertical motion and precipitation in the tropics (winter half-year mean, November–April). (a) Vertical pressure velocity (negative = upwards) in the PREIND experiment (colours and contours, $\text{Pa s}^{-1}$) with areas of strong precipitation (over $4 \text{mm d}^{-1}$) marked with ‘+’ symbols. (b) Difference in vertical pressure velocity between experiments MID-HOLOCENE and PREIND (colours, $\text{Pa s}^{-1}$) and vertical pressure velocity in experiment PREIND (contours at ±0.02 and 0.05 $\text{Pa s}^{-1}$, dashed for negative). (c) Difference in precipitation between experiments MID-HOLOCENE and PREIND (colours, mm d$^{-1}$) and precipitation in experiment PREIND (contours as marked, mm d$^{-1}$). Areas marked with crosses in (b,c) indicate statistical significance at the 95% level.

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stronger precipitation over the subtropical North Atlantic (from a very low baseline), extending into the Mediterranean; and
— weaker ascent (and precipitation) over the central and eastern tropical Pacific Ocean.

The broad character of these changes is akin to a La Niña-type situation in that ascent and precipitation is shifted westward in the tropical Pacific and over the maritime continent. This is consistent with various lines of palaeo-evidence indicating that the Mid-Holocene climate experienced fewer El Niño events than the Late Holocene (e.g. Clement et al. 2000; Stott et al. 2004; Wanner et al. 2008) and also the PMIP-2 models (Zheng et al. 2008).

Examining the tropical Atlantic sector (including South America and the western part of tropical Africa), however, reveals marked inconsistencies between the response seen in Vecchi & Soden (2007) and that shown in figure 6.

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Figure 7. Zonal wind and mid-tropospheric baroclinicity (winter half-year mean, November–April). (a) Zonal wind at 250 hPa (colours and contours, $\text{m s}^{-1}$) and divergent wind (arrows, $\text{m s}^{-1}$) in experiment PREIND. (b) Difference in zonal wind at 250 hPa (colours, $\text{m s}^{-1}$) and divergent wind (arrows, $\text{m s}^{-1}$) between experiments MID-HOLOCENE and PREIND, with zonal wind at 250 hPa in experiment PREIND (contours, $\text{m s}^{-1}$). (c) Eady growth rate (measured between 300 and 700 hPa; colours, $\text{s}^{-1}$) and zonal wind at 250 hPa in experiment PREIND. (d) Difference in Eady growth rate (measured between 300 and 700 hPa; colours, $\text{s}^{-1}$) between experiments MID-HOLOCENE and PREIND, with zonal wind at 250 hPa in experiment PREIND (contours, $\text{m s}^{-1}$). Areas marked with crosses in (b,d) indicate statistical significance at the 95% level. For (c,d) only data for the Northern Hemisphere are shown.
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Figure 8. The storm tracks in the TIMESLICE experiments; see appendix A for a detailed discussion of the precise diagnostics used. (a) Track density in experiment PREIND (colours, tracks in a 5° spherical cap per month, using relative vorticity maxima at 850 hPa). (b) Standard deviation of 2–6 day band-pass filtered meridional wind at 850 hPa in experiment PREIND (colours, m s\(^{-1}\)). (c) Difference in track density between experiments MID-HOLOCENE and PREIND (colours, tracks in a 5° spherical cap per month) and track density in experiment PREIND (contours at 8, 12, 16 tracks in a 5° spherical cap per month). (d) Difference in standard deviation of band-pass filtered meridional wind at 850 hPa between experiments MID-HOLOCENE and PREIND (colours, m s\(^{-1}\)) and the same field in experiment PREIND (contours at 3, 5 m s\(^{-1}\)). (e) As (c) but for experiment EARLY-HOLOCENE minus experiment 8 ka BP (colours) and experiment 8 ka BP (contours). (f) As (d) but experiment EARLY-HOLOCENE minus experiment 8 ka BP (colours) and 8 ka BP (contours). Grey areas in each plot indicate regions of high orography (over 1200 m) and areas marked with crosses in (c–f) indicate statistical significance at the 90% level. For (a,c,e) winter data from December to January are used; whereas for (b,d,f) data from the winter half-year November to April are used.

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particular, the ensemble-mean response to increasing greenhouse gases in GCMs studied by Vecchi & Soden (2007) shows weaker ascent over the land areas and stronger over the ocean—this is the same pattern (in sign as well as structure) as seen in figure 6b which is associated with reduced greenhouse gases.

This difference between the two responses suggests that the change in the greenhouse gas forcing is not playing the dominant role in the changes in the tropical ascent over the Atlantic sector in the TIMESLICE MID-HOLOCENE experiment. The change in the tropical ascent in this region is perhaps better understood as a response to insolation changes associated with orbital variations (figure 2).

In boreal winter, the Mid-Holocene tropics experience weaker insolation than that in the preindustrial experiment (approx. 20–30 W m\(^{-2}\), figure 2d,f) whereas in boreal summer the Mid-Holocene insolation is stronger than preindustrial (figure 2d). During boreal winter, these changes cause the land areas to become colder relative to the ocean, consistent with the differences in their surface heat capacities (i.e. land areas are cooler relative to the ocean in MID-HOLOCENE than in PREIND). This is consistent with weaker ascent over the land area (conceptually this is similar to the reverse of the process associated with the well-known northeastward extension of the West African monsoon system during boreal summer in the Mid-Holocene seen in the PMIP-2 models; Braconnot et al. 2007b), and the change is magnified by reduced latent heat release associated with less precipitation over the land areas of South America and tropical Africa (precipitation is reduced there in experiment MID-HOLOCENE compared with PREIND; figure 6c).

The geography of the African continent (the westward protrusion of Africa to the north of the equator) suggests that this process may contribute to an apparent southward shift in the boreal winter ITCZ over the tropical Atlantic (figure 6c). However, it should also be noted that the major ascent and precipitation regions during boreal winter over South America and tropical Africa occur well to the south of the equator and the oceanic boreal winter ascent maxima over the tropical Atlantic Ocean (figure 6a); in this sense, it could be argued that the weaker precipitation over land leads to a northward displacement of the boreal ITCZ within the Atlantic sector in MID-HOLOCENE. This, combined with the well-established northward shift in the boreal summer ITCZ (e.g. Braconnot et al. 2007b), is consistent with an annual-mean northward shift in the ITCZ in the Atlantic sector (cf. Haug et al. 2001).

The impact of changes in tropical ascent on the upper tropospheric subtropical jet in MID-HOLOCENE (compared with PREIND) is shown in figure 7. For much of the subtropical North Atlantic, the differences are rather small. However, over North Africa and East Asia there are marked differences in the upper tropospheric subtropical jet (figure 7b).

Over North Africa, the subtropical jet entrance is further south in MID-HOLOCENE than PREIND (by a few degrees latitude; figure 7b). This is associated with the modified ascent patterns in MID-HOLOCENE over tropical Africa and tropical North Atlantic (figure 6b). In particular, the southerly divergent wind in the upper troposphere over tropical Africa (approx. 20° E, 10° N in figure 7a) is stronger to the south of 5° N and weaker to the north (figure 7b), consistent with the subtropical jet over Africa being further south in MID-HOLOCENE than PREIND (figure 7b).
Near East Asia, the situation is rather different. Here the subtropical jet is stronger in MID-HOLOCENE than PREIND, although there is also some evidence of a slight southward shift into the north Pacific (100°E, 30°N in figure 7b). This is associated with the enhanced ascent over the Indian Ocean and western part of the maritime continent, which intensifies the upper tropospheric divergent flow to the north over southern Asia. It is also worth noting that the stronger upper tropospheric divergent flow in MID-HOLOCENE extends well over the Arabian Peninsula, consistent with stronger mid-tropospheric descent there (figure 6b; this feature will be returned to in §3).

The impact of these tropical circulation responses on the geographical structure of baroclinicity in the mid-troposphere (300–700 hPa) is shown in figure 7d, and the differences are clearly closely associated with the patterns of change in the subtropical jet structure (compare figure 7b,d). Over Southeast Asia (at approx. 30°N), the southward offset and intensification of the baroclinicity in MID-HOLOCENE is consistent with the expectation that the Northern Hemisphere winter storm tracks in the Mid-Holocene would be equatorward of those under preindustrial conditions (i.e. with a stronger but narrower Hadley cell from the zonal-mean arguments presented in §2a). A southward offset of the North Pacific storm track in MID-HOLOCENE relative to PREIND can indeed be seen in figure 8c,d (near 180°E, 30–40°N).

There is a slight suggestion of a similar southward offset in the North Atlantic (near 30°W, 35°N and stretching both east and west in figure 8c,d), suggesting that the tropical circulation changes may also be playing some role here (although it is unclear whether this is being driven by the tropical circulation changes in the western Atlantic area or other processes, e.g. land–sea contrast as discussed in §2c).

In contrast, the southward shift of the subtropical jet entrance over North Africa is associated with weaker baroclinicity at approximately 30°N (over the Mediterranean in the MID-HOLOCENE experiment baroclinicity is reduced by approx. 10% compared with PREIND; figure 7b,d). The weaker baroclinicity over the Mediterranean in MID-HOLOCENE suggests that that the changes in tropical circulation might act to produce a weaker storm track during the Mid-Holocene in the Mediterranean. Some indication of this can perhaps be seen over North Africa in the MID-HOLOCENE experiment (near 10°E, 25°N in figure 8c,d) but over the northern part of the Mediterranean and southern Europe the storm track appears slightly stronger in MID-HOLOCENE than PREIND. The Mediterranean storm track response is discussed in more detail in §3.

(c) The North American continent

The shape and characteristics of the North American continent have a strong influence on the strength and orientation of the downstream North Atlantic storm track. In particular, the extended meridional axis of the Rocky Mountains, embedded in the mid-latitude surface westerly flow, has been shown to generate an anticyclonic anomaly over the northern part of the mountain range and a cyclonic anomaly over the southern part (figure 9a; see Brayshaw et al. 2009 for further discussion of this). To the east of the mountain the surface anticyclonic anomaly leads to the southward advection of cold polar air towards
the continental coastline, whereas the surface cycloonic anomaly is associated with warm subtropical air moving poleward. The warm and cold temperature anomalies produced are clearly shown at 75°W, 20–40°N in figure 9a, and are associated with a strongly baroclinic region over the western part of the Atlantic basin (60–75°W, 35–40°N in figure 9b). Although part of this baroclinicity is attributable to the surface contrast at the coastline (cold land next to warm ocean), the large-scale barotropic circulation advecting warm and cold air masses (as described above) ensures that the signature of the baroclinicity penetrates well above the surface into the mid-troposphere (not shown).

The change in the North Atlantic storm track produced by ‘adding the Rocky Mountains’ is shown in figure 9c. The storm track has a stronger southwest–northeast tilt, consistent with the anomalous patterns of baroclinicity (figure 9b), and is associated with a southward shift in the storm track at the western side of the Atlantic (the storm track is strengthened at 90°W, 30°N by the order of 40%).

Consider now how these flow patterns may have been affected by palaeoclimate forcing during the Mid-Holocene. In the Mid-Holocene period, the winter insolation in the northern latitudes is lower than preindustrial levels whereas, in summer, it is greater than preindustrial levels (figure 2d). This will tend to increase the wintertime surface temperature contrast between the continent and the ocean, owing to their differing surface thermal capacities, as can be seen clearly in figure 10c (the difference between MID-HOLOCENE and PREIND surface temperatures is less negative over the ocean than the land). This is consistent with stronger baroclinicity near the eastern coastline of North America in MID-HOLOCENE compared with PREIND (figure 10d), although the change is reasonably small (approx. 5% at 70°W, 35°N). It should, however, be noted that the magnitude of the difference in winter insolation in the Mid-Holocene is less over the northern high latitudes than the northern mid-latitudes (figure 2f), and that this may act to reduce the temperature contrast near the coastline (the air advected towards the coastline from the north may be less cold relative to the mid-latitude air that it meets in MID-HOLOCENE than PREIND).

Consider now periods in the earlier Holocene, when the ice sheets formed in the last Ice Age remained over the northeastern part of the North American continent (figure 2b). These ice sheets would have been associated with a substantially larger and colder cold-pool of air in the northeast part of the continent (as seen in experiment EARLY-HOLOCENE; figure 10e). Assuming that the vertical height of the ice sheet is relatively small (it is prescribed to be less than 1km in the TIMESLICE simulations), the changes in the northwesterly flow over this region are likely to be similarly small. However, the same northwesterly flow would advect considerably colder air south and eastward towards and into the North Atlantic, promoting stronger baroclinicity over this region. Some evidence of this can, perhaps, be seen in figure 10f (near the east coast of North America and at 45°W, 45°N), although the overall change in baroclinicity is no greater than that seen in the MID-HOLOCENE experiment (approx. 5% of the background Eady growth rate). Directly around the southern limit of the North American ice sheet the baroclinicity is, however, much stronger (consistent with the surface temperature contrast between a permanent elevated ice sheet and land that is only partially snow-covered; figure 10f).

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Figure 9. The response to ‘adding the Rocky Mountains’ in an IDEALISED experiment (see appendix B for details). (a) Temperature anomalies at 925 hPa (shading, K) and streamfunction anomalies at 925 hPa (contours, at ±0.5, 1, 2, 3 × 10^6 m^2 s^-2, dashed for negative). (b) Eady growth rate anomalies at 925 hPa (shading, s^-1) and the full field Eady growth rate in the experiment without the Rocky Mountains (contours, at 5 and 8 × 10^-6 s^-1). (c) Anomalies in standard deviation of 2–6 day band-pass filtered meridional wind at 850 hPa (shading, m s^-1) and the full field in the experiment without the Rocky Mountains (contours, m s^-1). The black area indicates where the orography of the ‘Rocky Mountains’ exceeds 500 m. Areas marked with crosses in (c) indicate statistical significance at the 90% level.
Figure 10. Properties of the mid-latitude circulation in the TIMESLICE experiments for the winter half-year (November–April). (a) SAT (colours, °C) and streamfunction at 925 hPa (contours, interval $2 \times 10^6 \text{ m}^2 \text{s}^{-2}$) in experiment PREIND. (b) Eady growth rate at 925 hPa (colours, s$^{-1}$) in experiment PREIND. (c) Difference in SAT between experiments MID-HOLOCENE and PREIND (colours, °C), with streamfunction at 925 hPa in experiment MID-HOLOCENE (black contours) and PREIND (red contours). (d) Difference in Eady growth rate at 925 hPa between experiments MID-HOLOCENE and PREIND (colours, s$^{-1}$) and Eady growth rate in experiment PREIND (contours, interval $2 \times 10^{-6} \text{ s}^{-1}$). (e) As (c) but for experiment EARLY-HOLOCENE minus experiment 8 ka BP SAT (colours), EARLY-HOLOCENE streamfunction (black contours) and experiment 8 ka BP streamfunction (red contours). (f) As (d) but experiment EARLY-HOLOCENE minus experiment 8 ka BP (colours) and experiment 8 ka BP (contours). In each plot, grey areas indicate regions where surface pressure is less than 925 hPa, and in (c–f) areas marked with crosses indicate statistical significance at the 95% level. For (b,d,f) only data for the Northern Hemisphere are shown.
Overall, these differences suggest that the storm track along the North American coastline may have been stronger in the Early to Mid-Holocene than under preindustrial conditions, particularly during the Early Holocene period where large ice sheets remained in the northeast portion of the continent. Some evidence for this can be seen in the TIMESLICE integrations (figure 8c near 65° W, 40° N and figure 8e near 75° W, 35° N), but the model data are far from conclusive. Indeed, the enhanced baroclinicity and storm formation in the Early Holocene near 70° W, 30° N (figures 10f and 8f) may be more closely related to stronger warm southerly flow over the Gulf of Mexico (figure 10e), suggesting that the changes in the flow around the Rocky Mountains (linked to changes in the Pacific jet structure) may be as relevant to the strength of the storm track in the western North Atlantic as changes to the land–sea contrast at the east coast of North America.

Strong shallow eddy activity may have also occurred near the southern perimeter of the ice sheet (consistent with the strong surface baroclinicity there as seen in the EARLY-HOLOCENE experiment, figures 10f and 8e near 60° W, 45° N). However, the lack of atmospheric moisture associated with the cold, dry northerly surface flow (figure 10e) is likely to have opposed the development of strong, deep cyclogenesis in that region (no comparable increase in synoptic wind variance or storm intensity is seen around the ice sheet boundary; compare figures 8f and 11c with figure 8e). Further downstream of this, over the central North Atlantic where the cold dry continental air meets a supply of warm, moist air from the south (near 45° W, 40° N), there are signs of more intense storms (figures 8f and 11c) but with little change in their frequency (figure 8e).

(d) The North Atlantic Ocean

It is well established that mid-latitude SST gradients can have an impact on the mid-latitude storm track (e.g. Kushnir et al. 2002; Inatsu et al. 2003; Nakamura et al. 2004; Brayshaw et al. 2008). In general, it has been shown that the storm track is enhanced downstream of tight mid-latitude surface temperature gradients (Brayshaw et al. 2008).

Figure 12 shows the impact of an idealized dipole SST anomaly introduced into the North Atlantic in the Gulf Stream region (the anomaly substantially weakens the meridional SST gradient associated with the Gulf Stream at 40° N; figure 12a). The SST dipole leads to weaker near-surface baroclinicity over the Gulf Stream region (figure 12b), and a marked weakening of the storm track along a southwest–northeast axis across the whole of the North Atlantic basin (approx. 10% weaker in figure 12c). There are, however, suggestions of a weak intensification of the storm track near 0° E, 40–45° N, suggesting that the resultant storm track is more zonal in orientation.

The relevance of this to Holocene climate change is shown in figure 2e. At high latitudes, the annual mean insolation is greater in the Early Holocene than under preindustrial conditions (positive anomalies up to approx. +5 W m⁻² near the poles), but this decreases to negative anomalies in insolation at low latitudes (approx. −1 W m⁻² at 30° latitude). It is therefore reasonable to expect the high-latitude oceans to be warmer relative to the subtropical oceans in the Early to Mid-Holocene than under preindustrial conditions, thereby reducing the
Figure 11. Storm intensity in the TIMESLICE experiments. (a) Experiment PREIND. (b) Experiment MID-HOLOCENE minus PREIND. (c) Experiment EARLY-HOLOCENE minus 8ka BP. In (a,b), contours indicate intensities of 0.35 and $0.5 \times 10^{-5}$ s$^{-1}$ in experiment PREIND whereas in (c), the same contours are used but for experiment 8ka BP. Areas marked with crosses in (b,c) indicate statistical significance at the 90% level. The intensities shown in (a) are noticeably weaker than those seen in Hoskins & Hodges (2002), as they are calculated directly from the low-resolution spatially smoothed data which are used to detect cyclone tracks. However, comparing the relative magnitudes and patterns of change (e.g. (b)) with the full-field data (i.e. (a)) provides an indication of how storm intensities have changed across the model simulations. Winter data from December to January are used in each figure.

Examination of the MID-HOLOCENE simulation shows some evidence of these weaker meridional SST gradients during the winter half-year period (figure 10c). In general, the SAT difference (MID-HOLOCENE minus PREIND) is more negative at lower latitudes than higher latitudes over the Atlantic. This is particularly true in the Labrador Sea and immediately to the south of Greenland ($45^\circ$ W, $60^\circ$ N). The SAT structure in the northwest Atlantic therefore bears some resemblance to the idealized SST anomaly experiment in figure 12, and the reduction in the near-surface baroclinicity in figure 10d is consistent with that seen in the idealized experiment (figure 12b). From this change in baroclinicity, it might be reasonable to expect a weakened storm track in MID-HOLOCENE, particularly in the northern part of the North Atlantic basin (near the negative magnitude of the meridional SST gradient (note, however, that this assumes that there is little change in the ocean circulation and heat fluxes during this period, as discussed in §1b).

Examination of the MID-HOLOCENE simulation shows some evidence of these weaker meridional SST gradients during the winter half-year period (figure 10c). In general, the SAT difference (MID-HOLOCENE minus PREIND) is more negative at lower latitudes than higher latitudes over the Atlantic. This is particularly true in the Labrador Sea and immediately to the south of Greenland ($45^\circ$ W, $60^\circ$ N). The SAT structure in the northwest Atlantic therefore bears some resemblance to the idealized SST anomaly experiment in figure 12, and the reduction in the near-surface baroclinicity in figure 10d is consistent with that seen in the idealized experiment (figure 12b). From this change in baroclinicity, it might be reasonable to expect a weakened storm track in MID-HOLOCENE, particularly in the northern part of the North Atlantic basin (near the negative magnitude of the meridional SST gradient (note, however, that this assumes that there is little change in the ocean circulation and heat fluxes during this period, as discussed in §1b).
Figure 12. The response to ‘removing the Gulf Stream’ in an IDEALISED experiment (see appendix B for details). (a) SST anomalies (shading, °C) and full SST field in the experiment where the Gulf Stream is included (contours, °C). (b) Eady growth rate anomalies at 925 hPa (shading, s\(^{-1}\)) and the full field Eady growth rate in the experiment with the Gulf Stream (contours, at 5 and 8 \(\times 10^{-6}\) s\(^{-1}\)). (c) Anomalies in standard deviation of 2–6 day band-pass filtered meridional wind at 850 hPa (shading, m s\(^{-1}\)) and the full field in the experiment with the Gulf Stream (contours, m s\(^{-1}\)). The black area indicates where the orography of the ‘Rocky Mountains’ exceeds 500 m. Areas marked with crosses in (c) indicate statistical significance at the 90% level.
baroclinicity anomaly at 45° W, 50° N in figure 10d). A signal consistent with this is seen in figure 8d, which appears to be associated with a reduction in storm intensity rather than frequency (see figures 8c and 11b). The resulting storm track in MID-HOLOCENE appears to be more zonal across the North Atlantic (positive anomalies to the south of the main storm track axis in the northeast Atlantic; figure 8e), consistent with the response in the idealized SST experiment.

(e) North Atlantic storm track summary

The arguments presented in §2a–d suggest that the overall position of the North Atlantic storm track is likely to have been further south and weaker in the Early to Mid-Holocene compared with that under preindustrial conditions. In particular, this is consistent with the Early to Mid-Holocene experiencing (relative to preindustrial conditions)

— lower greenhouse gas concentrations and weaker insolation during boreal winter, creating narrower, stronger and possibly shallower Hadley cells;
— a reduced latitudinal SST gradient over the North Atlantic (associated with reduced baroclinicity over the ocean);
— a reduced latitudinal insolation gradient consistent with a reduced requirement for poleward energy from the subtropics to polar latitudes by the storm track; and
— reduced subtropical-to-mid-latitude SSTs, associated with reduced atmospheric moisture content and latent heat release during storm growth.

A southward offset and weakened North Atlantic storm track has been shown in the TIMESLICE experiments (figure 8c,d, e.g. the meridional dipole centred across 45° W, 45° N), with similar patterns of change in both storm frequency (track density; figure 8c) and storm intensity (figure 11b).

It should be noted that this interpretation is rather different from that expressed in Wanner et al. (2008), which is heavily based on an SST reconstruction derived from alkenone data presented by Rimbu et al. (2003). This SST reconstruction suggests that the mean boreal winter climate of the Mid-Holocene was characterized by an anomalously positive NAO (usually indicative of a stronger and more northerly storm track in the northeast Atlantic). In contrast to this, however, European palaeoclimate reconstructions based on lake-levels and pollen records indicate stronger winter westerly winds through southern Europe during the Mid-Holocene (e.g. Bonfils et al. 2004) suggesting a southward shift of the storm track (relative to preindustrial conditions) and a projection onto a negative NAO state during the Mid-Holocene. Further work is clearly required to understand this issue, particularly given the uncertainty surrounding Mid-Holocene climate simulations over the North Atlantic sector (Gladstone et al. 2005).

It has also been argued in this paper that owing to a combination of the insolation changes producing a stronger land–sea contrast in the Early to Mid-Holocene and the legacy of the ice sheets left remaining from the last Ice Age, stronger storm activity could be expected near the east coast of North America. This effect could extend into the mid-Atlantic, consistent with
the strong northwesterly flow from the northwest portion of the continent and the
southwesterly flow over the Gulf of Mexico (figures 10e and 8f). It is, however,
difficult to clearly separate this effect in the TIMESLICE simulations from the
baroclinicity changes associated with SST patterns in the North Atlantic. Indeed,
it is also difficult to identify this effect in the PMIP-2 Last Glacial Maximum
simulations which could be expected to have a stronger ice sheet response (Li &
Battisti 2008; Laine et al. 2009). The lack of a clear signal from the ice sheet
therefore suggests that either its effect is relatively small compared with the
other Holocene climate forcings, or that it may be underrepresented in the
model simulations.

3. The Mediterranean sector

As discussed in §1 and shown in figure 3b, the drivers of storm track changes
during the Holocene in the Mediterranean are complex. The evidence for
changes in each driver in figure 3b is reviewed below, using data from the
TIMESLICE experiments (shown in figures 13 and 14). The section concludes
with a brief summary.

(a) The strength and position of the North Atlantic storm track

Recent examinations of modern-day interannual variability of the Mediterra-
nean storm track indicate that a southward shift of the North Atlantic storm
track tends to be associated with stronger storm activity over the Mediterranean
(Trigo et al. 2000). In particular, as described in §1c, this is likely to be
associated with increased cyclogenesis as the synoptic scale flow propagating
eastward from the North Atlantic interacts with surface topography. In the MID-
HOLOCENE experiment, the stronger synoptic scale variability over southern
and central Europe is clearly visible in figure 13d, and is consistent with
stronger Mediterranean cyclogenesis in the major cyclogenesis areas (figure 14f),
particularly over the east coast of Spain and the Aegean (over the Gulf of Genoa
the signal is less clear). The enhanced cyclogenesis in these regions is furthermore
consistent with increased cyclone track density over the same areas (figure 13c)
and more widely across the northern part of the Mediterranean basin. It is,
however, difficult to detect any coherent and robust changes in Mediterranean
cyclone intensity (not shown).

It is therefore reasonable, based on the changes in the North Atlantic
storm track in isolation from any other climate responses, to expect that the
Mediterranean storm track would be stronger during the Early to Mid-Holocene
than under preindustrial conditions.

(b) Near-surface baroclinicity

The stronger seasonal cycle of insolation (figure 2d) during the Early to
Mid-Holocene tends to strengthen the wintertime surface temperature contrast
between land and ocean areas. Over the north coast of the Mediterranean, this
is associated with a stronger meridional temperature contrast (figure 14b shows
Figure 13. The Mediterranean storm track in the TIMESLICE experiments; see appendix A for a detailed discussion of the precise diagnostics used. (a) Track density in experiment PREIND (colours, tracks in a 5° spherical cap per month, using relative vorticity maxima at 850 hPa). (b) Standard deviation of 2–6 day band-pass filtered meridional wind at 850 hPa in experiment PREIND (colours, m s\(^{-1}\)). (c) Difference in track density between experiments MID-HOLOCENE and PREIND (colours, tracks in a 5° spherical cap per month) and track density in experiment PREIND (contours at 8, 12, 16 tracks in a 5° spherical cap per month). (d) Difference in standard deviation of band-pass filtered meridional wind at 850 hPa between experiments MID-HOLOCENE and PREIND (colours, m s\(^{-1}\)). (e) As (c) but for experiment EARLY-HOLOCENE minus experiment 8 ka BP (colours) and experiment 8 ka BP (contours). (f) As (d) but experiment EARLY-HOLOCENE minus experiment 8 ka BP (colours). Grey areas in each plot indicate regions of high orography (over 1200 m) and areas marked with crosses in (c–f) indicate statistical significance at the 90% level. For (a,c,e) winter data from December to January are used, whereas for (b,d,f) data from the winter half-year November to April are used.
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Figure 14. Storm track drivers and genesis in the Mediterranean in the TIMESLICE experiments. (a) SAT (colours, °C) and 925 hPa winds (arrows, m s⁻¹) in experiment PREIND. (b) Difference in SAT (colours, °C) and 925 hPa winds (arrows, m s⁻¹) between experiments MID-HOLOCENE and PREIND. (c) Eady growth rate at 925 hPa (colours, s⁻¹) and storm track genesis density (contours at 1.5 and 3 genesis events in a 5° spherical cap per month) in experiment PREIND. (d) Difference in Eady growth rate at 925 hPa (colours, s⁻¹) and storm track genesis density (contours at 1.5 and 3 genesis events in a 5° spherical cap per month) in experiment PREIND. (e) Storm track genesis density (colours, genesis events in a 5° spherical cap per month) in experiment PREIND. (f) Difference in storm track genesis density (colours, genesis events in a 5° spherical cap per month) and full-field storm track genesis density (contours at 1.5 and 3 genesis events in a 5° spherical cap per month) in experiment PREIND. Grey areas in (a–e) indicate regions of high orography (over 1200 m). Grey areas in (f) indicate areas of high orography and where the full-field genesis density is low (less than 1.2 genesis events in a 5° spherical cap per month). Areas marked with crosses indicate statistical significance at the 95% level (b), the 90% level (d) and the 70% level (f). For all figures winter data from December to January are used.
temperature differences that are more negative over land than oceans) and
stronger near-surface baroclinicity at the coast (figure 14d suggests differences
of the order of a few percent).

The increased baroclinicity in these areas is consistent with stronger storm
activity across the northern part of the Mediterranean basin. In particular, the
MID-HOLOCENE experiment (figure 14d,f) suggests consistency between the
changes in near-surface baroclinicity and cyclogenesis (compare the changes in
the two figures at 5° E, 42° N; 15° E, 42° N; and 22° E, 40° N, although the last of
these is less clear).

In isolation from all other climate responses, the changes in near-surface
baroclinicity therefore suggest that the Mediterranean storm track would be
stronger in the Early to Mid-Holocene than under preindustrial conditions.

(c) The subtropical jet entrance

As discussed in §2b, the subtropical jet entrance over western North Africa is
further south in the MID-HOLOCENE experiment than in PREIND (figure 7b),
consistent with a southward offset in the most baroclinic region in the mid-to-
upper troposphere (figure 7d). The impact such a change would have on the
Mediterranean storm track is far from clear.

In isolation, weaker mid-to-upper tropospheric baroclinicity could be expected
to weaken the storm track over the Mediterranean. However, such a reduction
in activity is not clear in the MID-HOLOCENE experiment (compare the blue
region over the Mediterranean in figure 7d with the storm track changes in
figure 13c,d). In contrast, the MID-HOLOCENE response suggests that the
MID-HOLOCENE storm track is weakened in the southern Mediterranean and
increased in the north, perhaps associated with the storm track becoming more
closely related to the surface temperature contrast at the northern coastline. Over
the Saharan cyclogenesis region (near 10° E, 23° N; figure 14f), there is evidence
of slightly stronger cyclogenesis to the south in the MID-HOLOCENE simulation.

(d) Mediterranean storm track summary

From the theoretical arguments presented above and the results presented
from the TIMESLICE simulations, it is therefore reasonable to expect that
the Mediterranean surface storm track of the Early to Mid-Holocene may have
been stronger than that seen under preindustrial conditions (in the TIMESLICE
simulations this change is associated with approx. 5% more storm tracks over the
northern part of the basin). However, it may also have been the case that the
storms in the storm track were shallower, consistent with weaker baroclinicity in
the middle and upper troposphere and also reduced latent heat release (associated
with the generally cooler winter temperatures). The relative strengths of the
genesis centres over the northern Mediterranean and the Sahara may also have
been different from those seen in the preindustrial climate.

Despite this, important aspects of these differences in the storm track remain
unclear. Three of these aspects are discussed below.

Firstly, under modern climate conditions, one usually expects that an increase
in storm activity would be associated with more precipitation. However, in the
generally cooler winter temperatures of the Early to Mid-Holocene, it is far
from clear from theoretical arguments alone whether precipitation would be
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Figure 15. Mediterranean winter precipitation (December–February) in the TIMESLICE experiments (mm d\(^{-1}\)). (a) Experiment PREIND. (b) Experiment MID-HOLOCENE minus PREIND. Areas marked with crosses in (b) indicate statistical significance at the 90% level.

stronger (because of more storm activity) or weaker (because of lower atmospheric moisture content). However, in the TIMESLICE simulations, the indications are that the EARLY-HOLOCENE and MID-HOLOCENE periods received more winter precipitation over the Mediterranean than PREIND (around 5–10% more over southwest Turkey; figure 15). This is consistent with many palaeorecords (e.g. Roberts et al. submitted) in terms of the sense of the change, but the magnitude of the change is generally rather less than that suggested by proxy data (e.g. Black et al. in press). Understanding this response is clearly an important issue as water supply is one of the most important factors for interpreting regional archaeology, and is crucial for the formation of many palaeo-proxy records.

Secondly, even if the Mediterranean storm track is ‘more active’ during the Early to Mid-Holocene, the propagation of these storms may be different. In particular, the TIMESLICE simulations suggest that during winter there is stronger mean descent over a region stretching from Southeast Asia towards the Eastern Mediterranean (figure 6b, associated with the changes in upper tropospheric divergence owing to the large-scale shift of the tropical ascent over the tropical Indian Ocean, figure 7b). This descent is likely to suppress precipitation over this region (extending well into the eastern part of the Fertile Crescent in the TIMESLICE simulations; figure 15b), and may deflect storms to the north (compare the track density differences in the Eastern Mediterranean in figure 13c with the near-surface winds in figure 14b; see also Enzel et al. 2003, 2008).

Thirdly, many palaeorecords, particularly but not solely in the Eastern Mediterranean, indicate a ‘shift’ in climate in the Mid-Holocene from ‘wetter’ to ‘drier’ (e.g. Roberts et al. (submitted) provide a recent review of the evidence in the Mediterranean while Steig (1999) highlights a change on a more global level). As discussed by Brayshaw et al. (in press a,b), the TIMESLICE simulations used

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here provide some suggestion of a similar change although, as noted there, it is difficult to provide a conclusive assessment of the model response owing to problems in the configuration of the TIMESLICE experiments (see appendix B for brief discussion of this). Nevertheless, as Brayshaw et al. (in press a,b) note, there remain intriguing suggestions that the strength of the Mediterranean storm track (and the precipitation associated with it) may change markedly in the model in response to the prescribed climate forcings around 6–10 ka BP (figs 13 and 14 of Brayshaw et al. in press a). This suggests that the combined forcing on Mediterranean climate system may have approached a form of threshold around this period, and that gradual changes in each of the climate forcings described here may have combined to produce a more rapid change in the Mediterranean hydroclimate. It is, however, necessary to approach this with some caution. Further simulations are required to test this hypothesis and, in particular, to confirm that the behaviour is not an artefact of the model configuration error (described in appendix B).

4. Summary and conclusions

This paper has identified and reviewed the principal drivers for Holocene climate change in the context of the North Atlantic and Mediterranean storm tracks. Developing a physically based understanding of these storm tracks and how they may have changed during the Holocene represents a small but essential step towards understanding past climate change in the region and its impact on the development of human civilization in the Mediterranean region.

Based on theoretical arguments and model simulations it is suggested that, during the Early to Mid-Holocene, the North Atlantic storm track was weaker and further south than its present position (or its position under preindustrial conditions). This is consistent with (during the Early Holocene)

— a stronger and narrower Hadley cell (associated with reduced atmospheric moisture content in the tropics);
— weaker meridional gradients of insolation (amongst other effects this leads to weaker SST gradients in the extratropics); and
— weaker insolation (particularly in winter) in the northern subtropics (associated with colder SSTs, reduced atmospheric moisture content and therefore latent heat release for storm growth).

Somewhat in contrast to this, storm activity may have been somewhat enhanced during the Early to Mid-Holocene along the coastline of North America. This enhancement is consistent with stronger land–sea contrast (associated with an enhanced seasonal cycle of insolation), although the evidence for an enhancement in the TIMESLICE simulations is inconclusive.

Over the Mediterranean sector, there is evidence to suggest that the near-surface storm track would have been rather stronger in the Early to Mid-Holocene than under preindustrial conditions. Such a situation is consistent with a more southward position of the North Atlantic storm track (triggering more Mediterranean cyclogenesis) and stronger land–sea contrast (creating stronger surface baroclinicity at the northern Mediterranean coastline). The Early to
Mid-Holocene Mediterranean storm track may, however, be shallower in the vertical (owing to reduced mid-tropospheric baroclinicity associated with a southward shift in the subtropical jet entrance over Africa) and individual storms less intense owing to reduced atmospheric moisture content (consistent with generally lower near-surface temperatures, although it should be noted that the TIMESLICE integrations do not suggest a particularly large change in winter SST over the Mediterranean). The penetration of the Mediterranean storm track into the Eastern Mediterranean may also have been affected by large-scale descent perhaps associated with changing patterns of tropical convection over the tropical Indian Ocean.

There remains much to be understood about climate change in the North Atlantic and the Mediterranean during the Holocene. The discrepancies between different model simulations (e.g. Gladstone et al. 2005) and the difficulty of interpreting and reconciling palaeo-observations (cf. Rimbu et al. 2003, 2004; Bonfils et al. 2004; Brewer et al. 2007; see §2e) emphasize the need for continuing dialogue between climate modelling, climate dynamical theory and palaeo-observations. It is hoped that this paper provides a small step towards this by identifying some of the key mechanisms involved in modifying the storm tracks during the Holocene, and how they link to large-scale climate properties and drivers in remote regions. The fact that such large-scale atmospheric linkages exist highlights the value of existing palaeoclimate syntheses combining modelling and palaeo-observations over large regions (such as Cheddadi et al. 1997; Jolly et al. 1998; Wanner et al. 2008; Wohlfahrt et al. 2008), and the need for continuing work in this area. Indeed, it is perhaps the large-scale climate reconstructions derived from proxy data that seem to represent the most practical targets for ‘testing’ or validating climate model simulations of the Early to Mid-Holocene (rather than attempting ad hoc site-by-site comparisons).

In order to better understand the Mediterranean hydroclimate, future work will focus on examining the basic properties of the Mediterranean storm track and its relationship to the Atlantic sector. This will also investigate the possibility that significant changes in the atmospheric circulation over the Mediterranean occurred between the Early and Mid-Holocene.

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Appendix A. Diagnostic techniques

Two analytical methods are used to diagnose storm track activity. The first approach is the standard ‘Eulerian’ method where 6-hourly model data are band-pass filtered (using a Lanczos filter; Duchon 1979) at each grid point to extract synoptic time scales (2–6 days), and variance maps of the resulting data are produced. The Lanczos filter used has a total width of 30 days and isolates the 2–6 day window well. In this paper, the field used is meridional wind at 850 hPa.
The second approach is a ‘Lagrangian’ feature tracking method developed by Hodges (1995) and described extensively in Hoskins & Hodges (2002). This method takes 6-hourly model data (relative vorticity at 850 hPa), removes the large-scale background flow, and tracks cyclonic circulation maxima. For a track to be recorded, a minimum lifetime of two days is required. Using this method, statistical information on the properties of individual storms can be collected (e.g. track density and storm intensity).

Throughout this paper, statistical significance is tested using the non-parametric Wilcoxon–Mann–Whitney (WMW) test (see Wilks 1995). For ‘Eulerian’ storm track statistics, data are resampled at 5 day intervals to remove serial correlations (see Brayshaw et al. (2009) for further explanation). For all other fields, it is assumed that data from each model year are uncorrelated.

Appendix B. Climate model configurations

To explore the potential changes in the storm tracks during the Holocene, two sets of model simulations are used. These both use a full GCM atmospheric model and are labelled the ‘IDEALISED’ and ‘TIMESLICE’ simulations. Full details describing the model configuration and validation for each can be found in Brayshaw et al. (2009) and Brayshaw et al. (in press a,b) respectively. As such, only a brief recap is provided here.

The ‘IDEALISED’ simulations use a moderately high-resolution version of the HadAM3 atmospheric model (1.875° × 1.25° × 30 vertical levels) with a fixed SST ocean (Pope et al. 2000). The model is configured to perpetual equinox conditions (i.e. the seasonal cycle is removed) with a simplified flat ‘3-continent’ world (North America, Eurasia, South America; see figure 9). SSTs are initially prescribed to be zonally symmetric, broadly resembling those seen in the winter North Atlantic (there is no sea ice in the model).

Two experiments are described in this present paper. In the first, ‘adding the Rocky Mountains’, an idealized Gaussian-bump version of the Rocky Mountains is placed over North America, shown by the black shaded area in figure 9. In the second, a more realistic SST structure is included over the North Atlantic (shown by the contours in figure 12a) and the impact of ‘removing the Gulf Stream’ (shown by the shading in figure 12a) examined. These experiments, while highly idealized, provide insight into the basic ingredients providing the large-scale forcing of the storm track relevant to Holocene climate change.

The ‘TIMESLICE’ simulations are twofold. Firstly, a standard-resolution version of the HadAM3 model (3.25° × 2.5° × 19 vertical levels) is coupled to a thermodynamic slab ocean (Hewitt et al. 2001), with the ocean heat transport prescribed to preindustrial levels. The output from the global model is then dynamically downscaled using a regional model, HadRM3, over the Mediterranean to approximately 50 km (the regional model is very similar to that used in the UK Met Office’s PRECIS system; http://precis.metoffice.com/docs/PRECIS_Handbook.pdf).

Several experiments are used to examine the atmospheric response to Holocene climate change. This includes changes in greenhouse gas concentrations (methane and carbon dioxide, see figure 2a), orbital variations (e.g. figure 2d), land ice sheets (figure 2b) and sea level. The prescribed changes in these forcings

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are similar (but not identical) to those used in the PMIP-2 intercomparison (Braconnot et al. 2007a), and are consistent with Jansen et al. (2007). Each simulation is at least 20 model years (after an initial spin-up of 5 years), and one ‘TIMESLICE’ simulation is carried out every 2000 years into the past from preindustrial conditions (approx. AD 1800, 280 ppm CO₂, before the large anthropogenic increase in CO₂). An additional simulation is carried out using present-day greenhouse gas concentrations (320 ppm CO₂, i.e. approximately that in the 1970s). The experiments are listed in table 1.

As discussed in the main text, it should be noted that some of the TIMESLICE experiments have been found to be affected by a subtle error in their configuration (affecting experiments 8, 10 and 12 ka BP only, and described in detail by Brayshaw et al. in press a,b), making it is difficult to use the TIMESLICE experiments as a simple time series stretching back to 12 ka BP. As such, a two-stage approach is adopted. Firstly, a time series stretching back to 8 ka BP is created using the ‘8 ka BP-NOICE’ experiment (which uses the orbital and greenhouse gas forcing relating to 8 ka BP, but with modern-day land ice sheets and sea level). The MID-HOLOCENE experiment is then defined as the average of experiments 8 ka BP-NOICE and 6 ka BP (see table 1). To examine the changes in the Early Holocene, we then use the EARLY-HOLOCENE experiment (the average of 10 ka BP and 12 ka BP) and compare this with experiment 8 ka BP: all three of these experiments are affected by the configuration error, thereby minimizing (although not removing) its influence. Further simulations are planned to repeat these Early Holocene integrations, but it is not believed that this configuration error significantly affects the discussion presented in this paper.

References


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