Composite Analysis of North Atlantic Extratropical Cyclones in NCEP–NCAR Reanalysis Data

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ABSTRACT

Composite analysis of North Atlantic midlatitudinal winter cyclones is performed using NCEP–NCAR reanalysis data for the 60-yr period from 1948 to 2007. The composites were developed using an advanced methodology involving the coordinate transform of cyclones into a nondimensional azimuthal coordinate system and the further collocation of fields. Composite analysis is performed for air–sea turbulent fluxes, heat content, precipitable water, and precipitation for 576 oceanic cyclones generated in the Gulf Stream area in winter (January–March) from 1948 to 2007. For the region of cyclone generation over the Gulf Stream, composites were analyzed for different cyclone intensities. Over the whole North Atlantic, composites were developed throughout the life cycle and for different cyclone types classified by the regions of their migration. These classifications allow the case-to-case variability to be minimized and the robustness of the composite to be boosted. In the region of cyclone generation over the Gulf Stream, characteristics of the composites strongly depend on the cyclone intensity quantified through the radial sea level pressure difference between the cyclone’s edge and its center. Stronger cyclone intensity implies larger turbulent fluxes in the rear of a cyclone and stronger precipitation in the forward part. Cyclones gradually dry with the water content and precipitation rate decreasing by about 40% and 50%–70%, respectively, during the lifetime. Although composites of air–sea turbulent fluxes show locally very strong positive fluxes in the rear part of the cyclone, the total air–sea turbulent fluxes provided by cyclones are not significantly different from the averaged background fluxes. This shows that the formation of extreme air–sea fluxes by cyclones is connected to the larger-scale circulation conditions, particularly to the cyclone–anticyclone transition zones.

1. Introduction

North Atlantic midlatitudinal cyclones play a key role in forming the synoptic variability of air–sea fluxes and transporting heat and moisture to Europe during the winter. Diagnostic and modeling studies show that cyclone activity controls the synoptic variability of surface turbulent fluxes over the Gulf Stream (Alexander and Scott 1997; Giordani and Caniaux 2001; Zolina and Gulev 2003; Shaman et al. 2010) and in the Kuroshio region (Bond and Cronin 2008; Taguchi et al. 2009). Statistical association of the North Atlantic synoptic activity with European climate variability was demonstrated in a number of earlier works (Rogers 1997; Gulev et al. 2002; Bojariu and Giorgi 2005). In this respect cyclone climatologies (Serreze et al. 1997; Blender et al. 1997; Sinclair and Watterson 1999; Sickmoller et al. 2000; Simmonds and Keay 2000a,b; Gulev et al. 2001; Hoskins and Hodges 2002; Hodges et al. 2003; Trigo 2006; Wang et al. 2006; Wernli and Schwierz 2006, Rudeva and Gulev 2007; Raible et al. 2008; Ulbrich et al. 2009, among others) provide good prospects for conventional statistical analysis of the role of the Atlantic cyclones in forming European climate anomalies. However, these analyses cannot answer two important questions: (i) to what extent do Atlantic cyclones provide the interface between air–sea fluxes and the heat–moisture transport to Europe, and (ii) which cyclones and by which mechanisms transport moisture to the European continent. In our work we approach these issues by performing a detailed quantitative study of the thermodynamic structure of cyclones.

For this purpose, the composite analysis of cyclone characteristics for specific regions and the stages of cyclone development look particularly attractive. Sinclair and Revell (2000) stressed that “compositing ... highlights...
basic common features while eliminating detail of individual events” and, thus, works “to improve the signal-to-noise ratio of diagnostic calculations by averaging out observational and diagnostic calculations by averaging out case-to-case variability.” Starting from the pioneering work of Petterssen et al. (1962), the composite analysis of cyclones was employed for many purposes using different methodologies. When the composites are based on a limited number of cases, a thorough selection reduces case-to-case variability and allows for a highly realistic representation of specific cyclone features. This approach helped to effectively characterize cyclogenesis (Sinclair and Revell 2000), to analyze explosively deepening cyclones (Sanders and Gyakum 1980; Lackmann et al. 1996, 1997) and rapidly decaying cyclones (Martin et al. 2001; Martin and Marsili 2002; McIay and Martin 2002), as well as the tropical to extratropical cyclone transitions in the Atlantic (Hart et al. 2006).

Alternatively, the composites based on a large sample can identify only some general features of cyclones because the case-to-case variability is quite large in this example. Bauer and Del Genio (2006) compiled several hundreds of cyclones identified in the reanalyses of the National Centers for Environmental Prediction and the National Center for Atmospheric Research (NCEP–NCAR) and the 40-yr European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis (ERA-40), as well as those from the Goddard Institute for Space Studies model simulations, and concluded that the model cyclones are shallower and drier compared to those in the reanalyses. Chang and Song (2006) used 2300 cyclones to produce seasonal composites of precipitation fields in the North Atlantic and the North Pacific and found the seasonal cycles of intracyclone precipitation to be significantly different in the two oceans. Field and Wood (2007) used about 1500 midlatitude cyclones to quantify the relationships between the cyclone strength and the liquid water path and cloud characteristics. Their work helped to advance our understanding of the patterns of intracyclone cloud organization reported upon earlier by Lau and Crane (1995) and Norris and Iacobellis (2005). An extension of this analysis to the validation of the cyclone water cycle in the NCAR Community Atmospheric Model performed in Field et al. (2008) argues for a new microphysics scheme to be implemented.

To find an optimal compromise between the size of the sample and the complexity of the composition methodology, for our purpose we needed to properly justify the selection of cases. North Atlantic oceanic cyclones are characterized by the similarity of their basic characteristics, especially for the rapidly intensifying cyclones (Sanders and Gyakum 1980; Rogers and Bosart 1986; Lupo et al. 1992; Wang and Rogers 2001; Yoshida and Asuma 2004). The similarity of an air–sea flux structure within the cyclones in the areas of cyclogenesis was found in some diagnostic studies (Alexander and Scott 1997; Zolina and Gulev 2003; Jacobs et al. 2005; Shaman et al. 2010) as well as in the model experiments of Roebber (1989), Chao (1992), and Giordani and Caniaux (2001). For the Southern Hemisphere the similarity of air–sea fluxes for different stages of cyclone development has been demonstrated by Yuan et al. (2009), who used the satellite-based tracks of Patoux et al. (2009). However, in the observational case studies the estimates of air–sea fluxes in cyclones vary by up to 100% (Bane and Osgood 1989; Yau and Jean 1989; Fosdick and Smith 1991; Neiman et al. 1993; Liu et al. 1997). Consistency in the cyclone size changes during their lifetimes was noted by Grotjahn and Castello (2000), Simmonds (2000), and Rudeva and Gulev (2007), although the quantitative estimates of the size increase during their deepening may vary from 30% to 150%. Ruprecht et al. (2002), Bauer and Del Genio (2006), and Chang and Song (2006) argue that cyclones play a leading role in moisture transport; however, they do not associate cyclone characteristics at different stages with this transport. Sorteberg and Walsh (2008) showed that moisture transport to the Arctic is more connected with the local processes rather than the Atlantic cyclones. The fact that many results agree qualitatively, but not quantitatively, makes it difficult to quantify the specific features of various cyclone types and to associate these types with air–sea exchange processes. In this work we strive to resolve these issues.

In this study we focus on the characteristics of the winter North Atlantic cyclones that are generated over the Gulf Stream and that develop over the North Atlantic. Our main goal is to quantify the structure of the air–sea turbulent heat fluxes, atmospheric moisture and heat content, and precipitation at different stages of cyclone evolution. This knowledge can provide further insights into the mechanisms transporting Atlantic heat and moisture to Europe. We start with descriptions of the tracking methodology (section 2) and the synoptic climatology of the selected cyclones (section 3). Section 4 describes the methodology of the development of the cyclone composites. The regional cyclone composites for the Gulf Stream area are presented in section 5. Section 6 shows the results of the composite analysis of the cyclone life cycle over the North Atlantic. In section 7 we present the evolution of the integrated components of the cyclone water balance. The role of the Atlantic midlatitudinal cyclones in forming averaged air–sea fluxes is considered in section 8. Finally, section 9 summarizes the results and discusses the role of the North Atlantic cyclones in moisture transport over the midlatitudes.
2. Cyclone identification and tracking

We have used NCEP–NCAR reanalysis data (Kalnay et al. 1996; Kistler et al. 2001) for the period 1948–2007. This is the output of an offline run of the T62 operational model in data assimilation mode. As the data assimilation input changed considerably during 1948–2007, especially since the beginning of the satellite era in 1979, the reanalysis data may have some time-dependent inhomogeneities, especially in the Southern Hemisphere (White 2000; Hines et al. 2000). However, in the Northern Hemisphere, the NCEP–NCAR data represent a reliable basis for the analysis of the natural variability over the last several decades. A detailed analysis of the role of different types of assimilated data in reanalyses was presented by Bengtsson et al. (2004).

For storm tracking we used sea level pressure (SLP) 6-hourly output on a 2.5° × 2.5° grid. The advantages and disadvantages of the use of SLP data for cyclone tracking have been discussed in numerous studies. The tracking schemes of Blender et al. (1997), Serreze et al. (1997), Geng and Sugi (2001), Pinto et al. (2005), Bauer and Del Genio (2006), and others use SLP for cyclone identification. However, many authors use vorticity for the tracking procedure (Sinclair 1997; Simmonds and Keay 2000a,b; Hoskins and Hodges 2002; Hodges et al. 2003). Vorticity allows for more accurate identification of the cyclone generation at stages earlier than those revealed by SLP. On the other hand, using vorticity requires the application of smoothing procedures that make results dependent on the selection of the smoothing parameters (Sinclair 1997). Comparisons of the tracking results based on SLP and vorticity can be found in Hodges et al. (2003), Rudeva and Gulev (2007), Raible et al. (2008), and in the review article of Ulbrich et al. (2009).

We used a numerical tracking algorithm developed at the Institute of Oceanology of the Russian Academy of Science (IORAS). This scheme was initially developed for SLP data but was later adopted for the vorticity tracking (Rudeva and Gulev 2007). This method is described in Jung et al. (2006), Rudeva and Gulev (2007), and Löptien et al. (2008). The scheme works on the polar stereographic projection (181 × 181 points, with the center at the North Pole) to which the original NCEP–NCAR SLP 2.5° × 2.5° data are interpolated. The tracking procedure starts with the dynamical interpolation of 6-hourly SLP fields onto hourly time steps discriminating between the step-by-step cyclone migrations and the distances between the centers of the neighboring cyclones (see, e.g., Jung et al. 2006; Rudeva and Gulev 2007). Then, the identification of cyclone centers is performed by searching for the local SLP minima (<1015 hPa) using three passes for 5-, 9- and 13-point windows. The tracking itself starts with the method of the closest neighbors (Murray and Simmonds 1991; König et al. 1993; Sinclair 1997; Hodges 1994) and then employs several passes of the analysis of cyclone propagation velocities ahead and backward, through which the crossing trajectories are sorted. Validation of the tracking scheme skills using the results of semimanual tracking (Gulev et al. 2001) shows a close resemblance to the reference dataset. The tracking output (coordinates, time, and corresponding central pressure) is used for computing grids of cyclone counts (Zolina and Gulev 2002) and parameters of the cyclone life cycle. We use the minimum SLP during the cyclone lifetime and the SLP radial difference between the cyclone’s edge (see section 4 for the definition) and its center to characterize the cyclone intensity, although Simmonds and Keay (2000a,b) and Lim and Simmonds (2002) use alternative metrics (the Laplacian of pressure or maximum wind speed). Cyclone mean and maximum deepening/filling rates (∆SLP and ∆SLPmax) were estimated from the adjacent 6-hourly central SLP values. Cyclone propagation velocity and the cyclone lifetime were estimated from the tracking output using a simple numerical procedure.

3. Synoptic climatology of some selected cyclones

We have analyzed the winter (January–March) cyclones generated over the area shown in Fig. 1 from 1948 to 2007. This region includes the Gulf Stream from its separation at Cape Hatteras to 50°W. We will refer hereinafter to this region as the Gulf Stream area, noting, however, that the Gulf Stream is also present in the regions outside of this area. This is one of the major storm formation areas, where the local diabatic heating preconditions cyclone generation and intensification (Yau and Jean 1989; Hoskins and Valdes 1990; Giordani and Caniaux 2001; Zolina and Gulev 2003). In the winter seasons, 30% of the total number of cyclones and over 35% of the deep cyclones (<980 hPa) were generated in this region. The trajectories qualified for further analysis only if the cyclone generation point had fallen within the blue box in Fig. 1a and no cyclones generated and developed farther upstream were included in the census. Additionally, we required that the cyclone lifetime ranged from 3 to 7 days and that the cyclone migration from the generation point exceeded at least 1000 km. Long-lived cyclones (>7 days) often become stationary systems and may experience regeneration while short-lived cyclones (<3 days) are less intense (75% of cyclones are shallower than 1000 hPa) and smaller in size by 15%–20% (figure not shown). Accounting for these systems produces additional uncertainties in the tracking and compositing. Out of 982 cyclones generated in this area, 406 (41%)
were excluded from our analysis. The final subset of 576 tracks selected for the analysis includes the typical oceanic cyclones generated over the ocean and characterized by faster and longer migrations compared to the continental cyclones (Gulev et al. 2001). The climatological distribution of cyclone counts (Fig. 1b) clearly identifies the North Atlantic winter storm track with the maximum number of tracks over the northwest midlatitudinal Atlantic. Figure 1c shows that the region of cyclone generation is clearly associated with the maximal SST gradients.

Figure 2 compares cyclone characteristics from our census with those for all winter cyclones identified over the Northern Hemisphere (NH) and the NH extratropical oceans during 1948–2007. About 11% of the selected cyclones are deeper than 960 hPa and more than 56% have a minimum central pressure lower than 980 hPa (Fig. 2a). For all NH cyclones and oceanic cyclones these estimates are 14% and 45%, respectively. According to Fig. 2b, 58% of the selected cyclones can be classified as rapidly intensifying storms with \( \langle \delta \text{SLP} \rangle_{\text{max}} \) exceeding 1 Bergeron (24 hPa per 24 h). This is 3 times the number of the rapidly deepening cyclones over the NH (about 20% for 1948–2007) and 2.5 times the number of these cyclones over the extratropical oceans. Roebber (1984, 1989) and Serreze (1995) recommend normalizing the deepening rates as \( \langle \delta \text{SLP} \rangle = \delta \text{SLP}(\sin \phi_{\text{ref}} / \sin \phi) \), where \( \phi_{\text{ref}} \) is the reference latitude of 42.5°N in order to account for the latitude dependence of planetary vorticity, while Sanders and Gyakum (1980) used normalization with respect to 60° latitude. This procedure inevitably shifts the histograms.

Fig. 1. (a) Tracks of the 576 selected cyclones generated over the Gulf Stream (blue line shows the area of generation) for the winter periods of 1948–2007. (b) Spatial distribution of cyclone counts for selected transients. Cyclone counts in Fig. 1b represent the number of cyclones for the whole period in 5° cells, normalized to the size of 5° cells at 45°N (~218,000 km²) [see Zolina and Gulev (2002) for details]. (c) Winter climatological SST over the North Atlantic and the regions of decay among four classes of cyclones.
Filtering of the stationary cyclones results in a higher cyclone propagation velocity compared to the mean velocity of the NH oceanic cyclones. A histogram of the propagation velocity of selected cyclones shows a peak at
~40 km h$^{-1}$ (Fig. 2c) with the contribution of the slowly propagating systems ($V < 30$ km h$^{-1}$) being twice as small in our census. Our census is characterized by a significantly more squeezed distribution of cyclone size compared to all NH oceanic cyclones. Averaged over the entire cyclone life span the cyclone radius in our census (750 km) is ~200 km larger than that for all NH oceanic cyclones (Fig. 2d) with about 50% of the cases captured in the range of 700–900 km. Similarly, the maximum effective radius (typically approached at the moment of the minimum SLP) is considerably larger in our census (Fig. 2e). The cyclone asymmetry ratio $\eta$ was estimated as the ratio between the smallest and the largest cyclone diameters with the smallest one lying within ±20° of the diameter orthogonal to the largest (Rudeva and Gulev 2007). Figure 2f shows that the selected cyclones are more symmetric. The modal $\eta$ values range from 0.5 to 0.7, being larger by 0.1–0.2 than for all the NH cyclones and oceanic cyclones.

Summarizing, our census is representative for the most intense and moderate North Atlantic winter cyclones generated over the Gulf Stream area and those storms that developed over the midlatitudinal ocean. Such characteristics of the selected cyclones as their larger size and smaller asymmetry presumably imply smaller uncertainties of the composites based on our census.

4. Development of cyclone composites

Rudeva and Gulev (2007) developed a methodology for estimating cyclone size and geometry based on the coordinate transform and collocation of the cyclone center (defined by the tracking scheme) with the center of the virtual polar coordinate system. The established 36-radii azimuthal grid makes the further interpolation and estimation of the cyclone dimensions computationally convenient. For each radius we found the locations where $\partial \text{SLP}/\partial r = 0$. These points identified the cyclone edge and formed the curve $M$, capturing the area $S_M$. The virtual radius of a circle of the same area (effective radius) was used as a measure of the cyclone size. In this work we extended this approach to the construction of composites. After the collocation of the cyclone centers from the ensemble (for a given region or a given moment during the life cycle), we applied the coordinate transform as in Rudeva and Gulev (2007) and established an azimuthal grid with a virtual pole in the cyclone center, 36 radii and a 50-km radial step, which was used for identification of the cyclone edge. Then, all fields were interpolated onto this grid using the method of local procedures (Akima 1970), which is based on a piecewise function with slopes at the junction points determined locally by a set of polynomials. If we compare Akima’s (1970) method with alternative procedures (e.g., spline functions), we see that it yields a higher degree of accuracy of interpolation and does not result in the unrealistic local extrema. At this stage the original cyclone shape is not yet modified. The next step was the transformation of the initial azimuthal grid into a normalized circular grid with 36 nondimensional radii $\rho$, spanning from 0 to 1 with 0.05 increments. Re-interpolation of all fields onto this new grid also carried by Akima’s (1970) method already transforms the initial cyclone form into a circle. Figure 3 shows in a simplified form the details of the two grids for one cyclone at the initial stage of its development during winter 1997. The original grid [36, 36] is equally spaced with 50-km time steps for all radii, while the normalized grid [36, 20] has different spaces for different radii. This procedure was applied to all 576 selected cyclones at every 6-hourly time step. The resulting fields in the polar coordinate system had unique dimensions of [36, 20] for 36 radii and 20 points along each radius and were used for further composition.
Ruprecht et al. (2002), Bauer and Del Genio (2006), Chang and Song (2006), and Field and Wood (2007) used a simpler approach when the cyclone centers were collocated, but no further normalization and transforms were applied. Field and Wood (2007), in their analysis of precipitation and cloud structure, argued against using the rotation and rescaling options, mentioning the difficulty of compositing data from reanalyses and satellites and the potential problems with the location of the frontal surfaces, which were rather important in their study. To show the importance of the scaling procedure, we derived the angular probability density function (PDF) of the actual cyclone radii along 36 directions for all 576 cyclones (Fig. 4a). In the rear part of the cyclone the distribution of the cyclone radii is clearly less scattered compared to the forward part. For 50% of the modal cyclones, the radii in the back part vary from 400 to 700 km, while in the forward part this range increases to 400–800 km. Figure 4b shows the circular distribution of the standard deviations (STDs) of the cyclone radii. At the rear of the cyclone the STDs are 13% smaller compared to those in the cyclone’s forward part. The scaling procedure is expected to account for stronger uncertainty in the composed fields in the cyclone’s forward part compared to its rear part.

Figure 5 shows the composites of SLP overplotted with the wind speed vectors designed according to our methodology, and a more simplified one, in which the radial normalization and the adjustment of the cyclone propagation vectors are not applied. These composites were designed for the region of cyclone generation and intensification, centered at 40°N, 60°W (Fig. 1a). To analyze the cyclone evolution at different stages, Rudeva and Gulev (2007) introduced a nondimensional cyclone lifetime by normalizing the cyclone age with the cyclone lifetime:

$$\tau_i = t_i / T,$$

where $$\tau_i$$ is the normalized cyclone age at the time moment $$i$$, $$t_i$$ is the actual age at the same moment, and $$T$$ is the cyclone lifetime. The actual cyclone age $$t_i$$ was defined as the time in hours passed from the generation point at which $$t_i = 0$$. This approach has recently been found to be effective and has been advanced by Schneidereit et al. (2010). The normalized age [(1)] in this area ranges from 0.1 to 0.3 (from 7 h to 2.1 days for our census). A traditional composite (Fig. 5a) is characterized by a minimum SLP of 997 hPa and values of 1010 hPa at the edge ($$\rho = 1$$). In Fig. 5b, SLP is reasonably represented by concentric circumferences, while in the composite in Fig. 4a the concentric structure of SLP holds just for the core. Simple collocation (Fig. 5a) tends to locate the wind speed maximum closer to the cyclone center because the averaging is performed for cyclones of different sizes. Figure 5b shows a more realistic localization of the wind speed maxima in the southwest sector at $$\rho = 0.5–0.6$$. Remarkably, STDs of the scalar wind speed (Figs. 5c and 5d) are 15%–20% larger in the simplified composite compared to ours. When the procedure of collocation for compositing is applied without the coordinate transform (Fig. 5c), the local maxima of STDs of the wind speed are observed in the northeast cyclone sector. Thus, the scaling procedure allows us to account for the large scatter in the cyclone size and for the inhomogeneous distribution of this scatter in the angular direction. Qualitatively, our composite of SLP and wind speed is quite consistent with
those presented by Bauer and Del Genio (2006), Chang and Song (2006), and Field et al. (2008). The sector captured by 180°–300° will be approximately associated with the cyclone’s rear part and the 0°–120° sector with the cyclone’s forward part.

The composites were built for turbulent sensible and latent heat fluxes (positive fluxes are directed from the ocean to the atmosphere), precipitation, precipitable water (PW), and heat content. Turbulent fluxes, precipitation and PW were taken as provided by the NCEP–NCAR reanalysis. In the reanalysis data, PW represents the total vertically integrated water vapor of an air column overlying a unit area of the earth’s surface. The heat content was computed through the integration of \( T \rho C_p \) (with temperature \( T \); air density \( \rho \); and air specific heat capacity at constant pressure \( C_p \)) within the column spanning from the surface to different geopotential heights between 850 and 50 hPa. Moore and Renfrew (2002) report that NCEP–NCAR turbulent heat fluxes are overestimated in the Gulf Stream area. The biases could be reduced by recomputing the fluxes from the reanalysis state variables, as in Gulev et al. (2007). However, for analyzing energy balances it was important to use the fluxes that were consistent with the other reanalysis variables.

Fig. 5. Composites of (a),(b) the SLP overplotted with the wind speed vectors and (c),(d) the corresponding STDs of the SLP and scalar wind speed, derived using (a),(c) a conventional approach and (b),(d) the methodology used in this study. SLP in (c),(d) is shown in black contours; wind speed is in color.
5. Cyclone composites in the early stage of cyclone development

First, we will analyze cyclone composites in the area of generation over the Gulf Stream (Fig. 1a). The spatial SST gradient (Fig. 1c) here is crucially important for forming strong surface fluxes under the influence of the southward advection. However, at the synoptic time scales (2–8 days) the SST gradient plays a passive role in steering the temporal variability of fluxes due to a very slow rate of change in the ocean compared to that in the atmosphere. Alternatively, the atmospheric dynamics over the SST front plays an active role in driving the temporal variability of the surface fluxes (Alexander and Scott 1997; Zolina and Gulev 2003; Shaman et al. 2010). Out of these two factors, the SST front determines where the locally high fluxes may occur, while the atmospheric dynamics determines when these extreme fluxes happen. In the cold season these two processes together form the low-level baroclinicity and largely determine the cyclone development over the Gulf Stream (Minobe et al. 2008; Yao et al. 2008; Nakamura and Yamane 2009; Hotta and Nakamura 2011). The intracyclone structure of surface latent turbulent flux (Figs. 6a and 6b) reveals the maximum of 300–320 W m$^{-2}$ in the cyclone’s rear part and the minimum (60–80 W m$^{-2}$) in the forward part. Maximum (>150 W m$^{-2}$) and minimum (<0 W m$^{-2}$) sensible heat fluxes are collocated with those of the latent heat flux. The STD of the turbulent fluxes between the composite members varies from 60%–80% of the mean values in the cyclone rear to about 120%–200% in the forward part. The heat content of the atmosphere below...
300 hPa (Fig. 6c) decreases in the south–north direction, with the sampling variability being about 2% of the mean value. Note that the absolute values of the STD of the heat content practically do not change with the increase in the column height from 850 to 50 hPa, which implies that the processes in the lower troposphere are almost totally responsible for the variability in the heat content. The maximum PW content of 27 kg m$^{-2}$ (Fig. 6d) is identified in the cyclone warm sector; it decreases threefold in the northern cyclone part where the relative sampling variability is, nevertheless, much higher compared to the location of the maximum. A precipitation composite (Fig. 6e) identifies a strongly localized maximum of 30–32 mm day$^{-1}$ in the cyclone’s forward part and a minimum of 4–6 mm day$^{-1}$ at the rear. STD values are slightly lower than the mean values in the location of the maximum in the forward part and significantly exceed them in the rear part of the cyclone.

High sampling variability of surface turbulent fluxes (Figs. 6a and 6b) results from the different cyclone intensities. We characterized cyclone intensities by the radial SLP difference between the cyclone’s edge and the center $G$. This definition is practically identical to the “cyclone depth” used by Simmonds and Keay (2009), with the only difference in the definition of the cyclone edge. Cyclone intensity in the Gulf Stream area (Fig. 7a) clearly decreases with growing $G$, with about 45% of cyclones characterized by $G < 8$ hPa. To select appropriate classes of cyclone intensity, we analyzed a percentile distribution (Fig. 7b) and selected four classes with approximately equal numbers of cases (<5, 5–10, 10–18, and >18 hPa). Figure 8 shows the cyclone composites for these classes of $G$ in the Gulf Stream area. Although qualitatively the structure is similar to that in Fig. 6, the quantitative differences between various cyclones are quite large. Maximum turbulent heat flux (sensible + latent, $Q_{he}$) in the rear of the intense cyclones ($G > 18$ hPa) reaches 700 W m$^{-2}$, falling in the forward part to the negative values. For the weak cyclones ($G < 5$ hPa), the maximum $Q_{he}$ in the rear part amounts to 300 W m$^{-2}$. Composites in Fig. 8 are characterized by smaller (compared to Fig. 6) case-by-case variabilities. Thus, in the rear part of the cyclone, the STDs vary from 100 to 120 W m$^{-2}$ for moderate and intense cases ($G > 5$ hPa), which constitutes 20%–45% of the mean values. For the weak cyclones the STD is, however, quite large (220–240 W m$^{-2}$ or 70%–80% of the mean values). Application of a $t$ test shows that surface flux differences between nearly all classes of cyclones in the rear parts are statistically significant at the 95% level (Table 1). Differences between the forward parts are only significant for very weak and very intense cyclones. Significance estimates in Table 1 were derived for the actual number of snapshots entered into the averaging, which is 4–6 times higher than the number of cyclones falling into those classes (shown in Fig. 8). The adjacent snapshots may be statistically dependent, which decreases the effective number of degrees of freedom. Nevertheless, estimates performed for the number of cyclones show that differences between maxima for most classes were statistically significant. With the
FIG. 8. Composites for the cyclones of different intensities in the Gulf Stream region: (a) sensible plus latent heat fluxes (W m\(^{-2}\)), (b) heat content (10\(^8\) J m\(^{-2}\)), (c) PW (kg m\(^{-2}\)), and (d) precipitation rate (10\(^{-5}\) kg m\(^{-2}\) s\(^{-1}\)). Dashed contours correspond to the STDs of the parameters. Vertical columns correspond to different cyclone intensities with the radial SLP difference \(G\) and the number of cyclones falling to each category shown in the top line.
increasing $G$, maximum fluxes in the rear increase drastically, while in the forward part they remain quite stable. This implies strengthening of the $Q_{be}$ contrast between the rear and the forward part with an increase in cyclone intensity. This feature also holds for the heat content, PW, and precipitation rate (Figs. 8b–d). The south–north differences in the heat content and in PW increase with $G$ by 4 and 2 times, respectively. Similarly, the spatial gradients of the precipitation rate are increasing with $G$ by about 2.5–3 times. Table 1 shows that the differences between heat content, PW, and precipitation rate in various cyclone classes are statistically significant at the 95% level in most cases.

In Fig. 9 we show the dependencies of the intracyclone differences between the maxima and the minima of the major cyclone characteristics (in the polar presentation of the composites) on the cyclone intensity $G$. Clearly, the larger the intensity, the stronger the dipole structure of the major cyclone characteristics, with correlations that exceed 0.75 and achieve 0.87 for turbulent fluxes. Over the SST front stronger winds in the rear part imply a sharp increase in turbulent fluxes (cold-air outbreak).

In the forward part under the same conditions the intensifying winds may imply stronger negative fluxes of sensible heat (warm-air outbreak). Thus, an increase in

### 6. Composites for different stages of the cyclone life cycle

#### a. Evolution of composites during the cyclone life cycle

Cyclones propagating over the North Atlantic (Fig. 1) decay in various regions. This potentially implies a large spread of cyclone characteristics even for the same stages of development due to strong regional differences in the background conditions of both the ocean and the lower atmosphere. The larger spread will be associated with the latest stages of the cyclone development. Due to varying cyclone lifetimes (from 3 to 7 days in our census), the direct collocation of cyclones at different times during their life cycles is rather difficult. To build the composites, we used the normalized cyclone age $\tau$ [(1)], which varies from 0 to 1 for all cyclones and, thus, allows for comparison of cyclones with different lifetimes. In the next step cyclone characteristics were interpolated onto regular time steps $\Delta \tau = 0.05$ for which cyclone composites were derived. In all other respects the compositing methodology was as described in section 4.

Figure 10 shows the evolution of the composites of the SLP overlaid with wind vectors. At the moment of generation the central pressure is still relatively high (1005 hPa) and the radial SLP differences are rather

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<td>$5–10$</td>
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<td>$-0.01$</td>
<td>$H$</td>
</tr>
<tr>
<td>$-1.83^a$</td>
<td>PW</td>
</tr>
<tr>
<td>$-1.38^a$</td>
<td>$P$</td>
</tr>
<tr>
<td>$-44.07^a$</td>
<td>$-24.84^c$</td>
</tr>
<tr>
<td>$-0.09^b$</td>
<td>$-0.08^b$</td>
</tr>
<tr>
<td>$-3.35^a$</td>
<td>$-1.52^a$</td>
</tr>
<tr>
<td>$-1.94^a$</td>
<td>$-0.56^a$</td>
</tr>
<tr>
<td>$-119.51^a$</td>
<td>$-100.28^a$</td>
</tr>
<tr>
<td>$-0.32^a$</td>
<td>$-0.31^a$</td>
</tr>
<tr>
<td>$-6.37^a$</td>
<td>$-4.54^a$</td>
</tr>
<tr>
<td>$-2.72^a$</td>
<td>$-1.34^a$</td>
</tr>
</tbody>
</table>

| $G$ | $<$1% | 90% level | 95% level | 99% level |
|----------------------------------------|----------------------------------------|
| $Q_{be}$ | $3.14$ | $4.13$ | $5.22$ |
| $H$ | $0.23$ | $0.29$ | $0.35$ |
| PW | $0.35$ | $0.42$ | $0.49$ |
| $P$ | $1.04$ | $1.16$ | $1.28$ |

**Table 1. Differences in turbulent heat fluxes $(Q_{be}, W m^{-2})$, heat content $(H, 10^2 J m^{-2})$, PW $(kg m^{-2})$, and precipitation rate $(P, 10^{-3} kg m^{-2} s^{-1})$ between cyclones of different intensities and estimates of their statistical significance ($\tau$ test). Differences between the parameters in the locations of the maxima and minima are given above and below the main diagonal, respectively. Values for smaller intensities are subtracted from those for larger intensities.**
small (Fig. 10a). At $\tau = 0.2$, the mean $G$ increases to 16 hPa (Fig. 10b) and reaches the maximum of 20 hPa at $\tau = 0.5$ (Fig. 10c). Although the SLP at the cyclone edge decreases by $\sim 10$ hPa (likely due to the changes in the background SLP), the central pressure at $\tau = 0.5$ decreases strongly to 990 hPa. This results in increasing $G$ by about 4 times during the cyclone deepening stage. Wind speed composites (Fig. 10) reveal the increase in the wind speed maximum (southward of the cyclone center): from 8–9 m s$^{-1}$ at the moment of generation to 14–16 m s$^{-1}$ at $\tau = 0.5$. At the cyclone decay stage, the wind speed in the southwest part of the cyclone weakens to less than 10 m s$^{-1}$ with the local maximum shifted to the cyclone margin.

Changes in air–sea heat fluxes along the trajectory might have a background component driven by the climatological distribution of the fluxes (Zolina and Gulev 2003). Winter differences in $Q_e$ between the Gulf Stream region and the cyclone decay area (e.g., the northeast Atlantic) may amount to more than 100 W m$^{-2}$. On the other hand, cyclones, while developing, also change their characteristics (Fig. 10). It is difficult to discriminate the
impacts of the background and of the changes in cyclone characteristics, since these are synoptic systems that largely form the mean fields. Keeping this in mind, we show in Figs. 11a and 11b the composites of latent and sensible heat fluxes for different times of the normalized cyclone lifetime for our census. During the early stage of development, turbulent fluxes sharply grow in the cyclone’s back part, approaching the maxima $300 \text{ W m}^{-2}$ for $Q_e$ and $150 \text{ W m}^{-2}$ for $Q_h$ at $t = 0.1–0.2$, when the composites become very close to those in Fig. 6. The values of $Q_h$ and $Q_e$ in Fig. 11 may be somewhat smaller compared to those in Fig. 6, since the range $0 < \tau < 0.2$ spreads over a larger area than the region of cyclone generation (Fig. 1) expanding also to the middle North Atlantic. At $\tau \approx 0.5$, turbulent fluxes at the cyclone’s rear decrease to $220 \text{ W m}^{-2}$ for $Q_e$ and $120 \text{ W m}^{-2}$ for $Q_h$. During the cyclone’s decay stage ($\tau = 0.9$), both $Q_h$ and $Q_e$ in the cyclone’s rear part become quite small (about $120 \text{ W m}^{-2}$ for $Q_e$ and $40–60 \text{ W m}^{-2}$ for $Q_h$). During the cyclone’s life cycle the gradient of turbulent heat fluxes between the rear and the forward cyclone parts decreases sharply from $250 \text{ W m}^{-2}$ at $\tau = 0.2$ to about $40–60 \text{ W m}^{-2}$ at $\tau = 0.9$ for $Q_e$. Although the case-by-case variability of surface turbulent fluxes (Fig. 10) is quite large (STDs are typically $80%–100\%$ of the mean values), Table 2 shows that at least in the rear part the differences between the different time moments (Fig. 11) are statistically significant at the $95\%$ level according to a Student’s $t$ test.

Composites of the cyclone heat and water vapor contents (Figs. 11c and 11d) demonstrate how the combined effects of the northward displacement and changing cyclone intensity steer the cold- and warm-air advection in the cyclone area. The effects of the cold-air advection associated with the formation of the cold tongue ($H < 2.5 \times 10^9 \text{ J m}^{-2}$) are clearly seen at $\tau = 0.5$. Figure 11d reveals the strongest intracyclone gradient of PW (from $9$ to $27 \text{ kg m}^{-2}$) at $\tau = 0.1$ associated with the maximum PW in the southeast sector. At the later stages composites show gradual cyclone drying, with the PW decreasing by $60\%$ during the lifetime and a PW maximum in the southeast part of a cyclone. Although the warm conveyor belt is located westward of the PW maximum ahead of a cold front (Harrold 1973; Browning 1999; Carlson 1998), we note here that PW in Fig. 11d is integrated over the column, which can explain the eastward shift of the location of the PW maximum. STDs of the PW at all stages of cyclone development range from $6$ to $8 \text{ kg m}^{-2}$ (about $20\%–40\%$ of the mean values and less than $50\%–60\%$ of the intracyclone PW gradient at $\tau = 0.1–0.2$), clearly indicating that the differences between the composites at different stages of a cyclone’s lifetime are statistically significant (Table 2). Figure 11e shows that strong precipitation of more than $30 \times 10^{-5} \text{ kg m}^{-2} \text{ s}^{-1}$ (26 mm day$^{-1}$) in the cyclone’s forward part at $t = 0.1–0.2$ is halved at $t = 0.5$ and reduced by $\approx 80\%$ during the cyclone’s decay stage. Although the sampling variability of precipitation is close to the mean values, the changes between $\tau = 0.2$ and $0.5$ and between $\tau = 0.5$ and $0.9$ are statistically significant at the $95\%$ level (Table 2). In the rear part of the cyclone, the precipitation experiences a much smaller decrease through the life cycle (from $5–6 \times 10^{-5}$ to $3–4 \times 10^{-5} \text{ kg m}^{-2} \text{ s}^{-1}$, or approximately from $5$ to $3 \text{ mm day}^{-1}$) implying a weakening of the intracyclone precipitation contrasts from $25–30 \times 10^{-5} \text{ kg m}^{-2} \text{ s}^{-1}$ at $\tau = 0.1–0.2$ to about $5 \times 10^{-5} \text{ kg m}^{-2} \text{ s}^{-1}$ at $\tau = 0.9$. The evolution of precipitation (Fig. 11) is consistent with the “drying” of the cyclone core over the midlatitudinal ocean.
FIG. 11. Cyclone composites at different moments of the normalized lifetime: (a) latent heat flux (W m$^{-2}$), (b) sensible heat flux (W m$^{-2}$), (c) heat content ($10^8$ J m$^{-2}$), (d) PW (kg m$^{-2}$), and (e) precipitation rate ($10^{-5}$ kg m$^{-2}$ s$^{-1}$). Dashed contours correspond to the STDs of the parameters. Vertical columns correspond to different moments of the normalized cyclone life cycle, which are shown in the top line.
TABLE 2. Differences in latent heat flux ($Q_e$, W m$^{-2}$), heat content ($H$, $10^8$ × J m$^{-2}$), PW (kg m$^{-2}$), and precipitation rate ($P$, $10^{-4}$ kg m$^{-2}$ s$^{-1}$) between cyclones at different stages of their development ($\tau$) and estimates of their statistical significance ($t$-test). Differences between the parameters in the locations of the maxima and minima are given above and below the main diagonal, respectively. Values for smaller $\tau$ are subtracted from those for larger $\tau$.

<table>
<thead>
<tr>
<th>Normalized cyclone age, $\tau$</th>
<th>Normalized cyclone age, $\tau$</th>
<th>Normalized cyclone age, $\tau$</th>
</tr>
</thead>
<tbody>
<tr>
<td>0</td>
<td>$Q_e$</td>
<td>$Q_h$</td>
</tr>
<tr>
<td>0.2</td>
<td>-50.40a</td>
<td>28.29a</td>
</tr>
<tr>
<td>0.5</td>
<td>-43.10a</td>
<td>78.69a</td>
</tr>
<tr>
<td>0.9</td>
<td>-54.64a</td>
<td>11.54a</td>
</tr>
<tr>
<td></td>
<td>-29.89a</td>
<td>8.73a</td>
</tr>
<tr>
<td></td>
<td>-0.74a</td>
<td>-0.33a</td>
</tr>
<tr>
<td></td>
<td>-5.04a</td>
<td>-0.90a</td>
</tr>
<tr>
<td></td>
<td>-2.40a</td>
<td>-0.37</td>
</tr>
<tr>
<td></td>
<td>-58.73a</td>
<td>-15.63a</td>
</tr>
<tr>
<td></td>
<td>-21.71a</td>
<td>16.91a</td>
</tr>
<tr>
<td></td>
<td>-0.76a</td>
<td>-0.35a</td>
</tr>
<tr>
<td></td>
<td>-4.52a</td>
<td>-0.38</td>
</tr>
<tr>
<td></td>
<td>-2.62a</td>
<td>-0.59a</td>
</tr>
</tbody>
</table>

| $Q_h$ | $H$ | PW | $P$ |
| 0.16a | 0.64a | 1.14a |
| 0.27a | 6.37a | 14.22a |
| -2.57b | 12.21a | 22.18a |
| -4.14a | 6.10a | 13.95a |
| -2.03a | 14.78a | 24.75a |
| -54.64a | -11.54a | 105.9a |
| -29.89a | 8.73a | 66.75a |
| -0.74a | -0.33a | 0.50a |
| -5.04a | -0.90a | 7.85a |
| -2.40a | -0.37 | 9.97a |
| -58.73a | -15.63a | -4.09a |
| -21.71a | 16.91a | 8.18a |
| -0.76a | -0.35a | -0.02 |
| -4.52a | -0.38 | 0.52b |
| -2.62a | -0.59a | -0.22 |

a Significant at the 99% level.
b Significant at the 95% level.
c Significant at the 90% level.

b. Lifetime composites for different cyclone types

For further analysis we split all storms into four classes (Fig. 12a) grouped by the region of cyclone decay: cyclones propagating to the Labrador Sea and Baffin Bay north of 60°N (LAB), cyclones reaching the Norwegian and Greenland Seas east of 25°W (NE), cyclones decaying in the central Atlantic from 45° to 60°N (MID), and those propagating to the region 30°–45°N, 25°W–10°E (SE). These decay regions are also shown in Fig. 1c. In total, 277 cyclones out of 576 fall into these categories. Their typical lifetime is 4–4.5 days and they move over 2000 km from their point of generation. The remaining 299 cyclones have shorter lifetimes (3–3.5 days), shorter migration distances, and they decay either southwest of the MID region or over the continental Canada. The composites of 500-hPa height at the initial stage of the life cycle (Figs. 12b–e) associate cyclone classes with characteristics of the mean flow preconditioning their propagation to different regions. These composites clearly imply the positive NAO pattern for the NE cyclones and the negative NAO-like pattern for the MID cyclones. The LAB and SE cyclones are associated with anticyclonic and cyclonic blocking conditions, respectively. Table 3 shows that the LAB and NE cyclones are most intense and faster deepening. About 75% of NE cyclones and 56% of LAB cyclones reach the central pressure of <980 hPa. The SE cyclones are typically weaker (11% reach 980 hPa) and slowly deepening. These cyclones are embedded in a higher absolute background SLP and this affects their central pressures to a larger extent than those of, for example, the LAB cyclones. The NE cyclones exhibit the strongest $Q_{he}$ (up to 470 W m$^{-2}$) and the strongest intracyclone gradient of fluxes. At $\tau = 0.5$, the MID and NE cyclones are characterized by the strongest $Q_{he}$ with the maxima at 400–420 W m$^{-2}$ and 360–370 W m$^{-2}$, respectively. For the SE cyclones, the minimum $Q_{he}$ values are smaller (220–230 W m$^{-2}$) and not strongly localized compared to the MID and NE cyclones. For the cyclone decay stage ($\tau = 0.9$) the turbulent fluxes in the rear part become considerably weaker and range from 100–140 W m$^{-2}$ for the LAB cyclones to 200–260 W m$^{-2}$ for the MID cyclones. Maximum turbulent fluxes in the cyclone’s forward part at $\tau = 0.9$ range from $-40$ W m$^{-2}$ for the LAB cyclones to $+40$ W m$^{-2}$ for the MID and NE cyclones. The Labrador Sea is known to have extremely high turbulent fluxes in winter, especially if the fluxes are taken from the NCEP–NCAR reanalysis (Moore and Renfrew 2002). However, these extreme fluxes are associated with the cyclones generated in this area and with the cyclones propagating south of Greenland (MID and NE), rather than with the cyclones decaying over the Labrador Sea. Thus, it is not surprising that at $\tau = 0.5$ and 0.9 the turbulent fluxes in the LAB cyclones are smaller than those for the MID and NE cyclones. The case-by-case variability for the composites in Fig. 13 is considerably smaller compared to Fig. 11, with STDs being 45%–65% of the mean values.

The overall structure of the heat content (below 300 hPa) for selected cyclone types (figure not shown) is in agreement with Fig. 11 and is characterized by the low sampling variability. For the LAB, MID, and NE cyclones, the heat content reasonably decreases over the cyclone...
life cycle as these cyclones propagate northward. For the SE cyclones propagating zonally in the relatively homogeneous background, changes in the heat content are much smaller. The strongest PW changes (figure not shown) are identified for the LAB cyclones (a 5–6 times decrease during the lifetime). This is consistent with the propagation of the LAB cyclones into the colder and drier background compared to the other cyclone types. Weaker changes (2–3 times) are observed for the MID and NE cyclones and the smallest decrease of 15%–20% is identified for the SE cyclones. Differences between PW maxima and minima drop by 3–4 times during the life cycle with the strongest homogenization of PW for the LAB and NE cyclones. Table 4 shows relative PW changes with respect to the PW maximum at every 0.1 step of the normalized lifetime. For the LAB and MID

Table 3. Major characteristics of the cyclone life cycles for the four subsets of trajectories shown in Fig. 12 (see text for definitions). First number in each cell indicates mean, and number in parentheses stands for STDs.

<table>
<thead>
<tr>
<th>Cyclone category</th>
<th>LAB (55)</th>
<th>NE (78)</th>
<th>MID (112)</th>
<th>SE (28)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Min SLP during the life cycle (hPa)</td>
<td>970 (12)</td>
<td>962 (13)</td>
<td>973 (14)</td>
<td>988 (12)</td>
</tr>
<tr>
<td>Radial SLP diff at the moment of SLP_{min} (hPa)</td>
<td>29 (12)</td>
<td>29 (14)</td>
<td>24 (12)</td>
<td>14 (10)</td>
</tr>
<tr>
<td>Radial SLP diff at τ = 0.5 (hPa)</td>
<td>21 (13)</td>
<td>21 (12)</td>
<td>21 (12)</td>
<td>10 (6)</td>
</tr>
<tr>
<td>Cyclone lifetime (days)</td>
<td>4.75 (1.0)</td>
<td>5.0 (1.0)</td>
<td>4.5 (1.0)</td>
<td>5.25 (1.0)</td>
</tr>
<tr>
<td>Mean deepening rate [hPa (6 h)^{-1}]</td>
<td>-4.0 (2.2)</td>
<td>-3.7 (1.8)</td>
<td>-3.4 (2.2)</td>
<td>-1.9 (1.5)</td>
</tr>
<tr>
<td>Max deepening rate [hPa (6 h)^{-1}]</td>
<td>-7.7 (3.1)</td>
<td>-7.8 (2.7)</td>
<td>-7.1 (3.1)</td>
<td>-4.7 (2.6)</td>
</tr>
<tr>
<td>Cyclone propagation velocity (mean, km h^{-1})</td>
<td>44 (11)</td>
<td>51 (9)</td>
<td>44 (10)</td>
<td>43 (10)</td>
</tr>
<tr>
<td>Effective cyclone radius at τ = 0.5 (km)</td>
<td>854 (255)</td>
<td>769 (201)</td>
<td>805 (324)</td>
<td>620 (520)</td>
</tr>
</tbody>
</table>
cyclones the most intense drying occurs at $\tau = 0.3-0.7$, that is, approximately immediately after the moment of the maximum PW. At the same time, for the NE and SE cyclones for which the maximum PW is achieved at a later stage ($\tau = 0.3-0.4$), the most intense drying is observed at the latest stages of the cyclone development ($\tau = 0.6-1.0$). Precipitation composites for the individual cyclone types are shown in Fig. 13b. At $\tau = 0.5$, the maximum precipitation in the cyclone forward part decreases by more than 2 times and further drops at $\tau = 0.9$ by 4–5 times for all classes. Precipitation in the cyclone’s rear part varies from $2 \times 10^{-5}$ kg m$^{-2}$ s$^{-1}$ and shows little change during the lifetime. STDs of the parameters in the composites for individual cyclone classes are 20%–40% smaller compared to the averaged composites. In the forward part, the STDs range from 60% to 80% of the means, and the differences between the values at different moments of the cyclone life cycle are statistically significant.

7. Evolution of the integrated components of the cyclone water balance

Composite analysis also allows us to estimate the integrated components of the cyclone water vapor balance during its life cycle. Integration of precipitation $P$ and surface evaporation $E$ within the cyclone in the polar coordinates $[r, \theta]$ was provided as

$$
\langle x \rangle = \int_{S} x \, ds = \int_{S} x \, r \, dr \, d\theta,
$$

Fig. 13. Cyclone composites at different stages of cyclone development for different cyclone types: (a) latent plus sensible turbulent flux ($W$ m$^{-2}$) and (b) precipitation rate ($10^{-5}$ kg m$^{-2}$ s$^{-1}$). Dashed contours correspond to the STDs of the parameters. Vertical columns correspond to different cyclone types indicated in the top line along with the numbers of cyclones falling to each category.
where $x$ denotes the parameter ($E$ or $P$) and $\langle x \rangle$ is the value integrated over the cyclone area. For $PW$ the integrated value also includes vertical integration over the height of atmospheric column $h$:

$$\langle x \rangle = \int \int_S x \, ds \, dh = \int \int_S x r \, dr \, d\theta \, dh.$$  

(3)

At every time step the water vapor balance of the cyclone can be described as

$$\frac{\partial PW}{\partial t} + B(\text{conv} + \text{err}) = \langle E \rangle - \langle P \rangle,$$

(4)

where $E$ and $P$ are the surface evaporation and precipitation, respectively; $\partial PW/\partial t$ represents the temporal change of the total water content; and the angular brackets indicate integration over the cyclone domain according to (2) and (3). The term $B$ on the lhs of (4) can be attributed to the moisture convergence due to lateral advection of water vapor into the cyclone domain (conv) and to the uncertainties (err). The coordinate transform did not allow for the accurate direct computation of the convergence term and its estimation was achieved by the residual approach. This is an extension of the atmospheric water recycling concept (Brubaker et al. 1993; Trenberth 1999; Trenberth et al. 2003) to the analysis of a moving domain (cyclone) for which we also estimate a temporal change in the water content. The case-by-case variability of surface fluxes and precipitation at the development stage is generally lower compared to the means, while at the decay stage STDs frequently exceed the means (Figs. 11 and 13). Thus, the estimates of different terms in (4) were derived for the initial stage of the life cycle ($0 < \tau < 0.2$), mid-deepening range ($0.2 < \tau < 0.4$), and the stage of maximum development ($0.4 < \tau < 0.6$).
Figure 14a shows estimates for cyclones of different intensities quantified for this analysis as the maximum radial SLP difference during the lifetime (typically achieved at $\tau = 0.5$), and not as in Figs. 8 and 9 where it was estimated for the initial cyclone stage. For all cyclones in the Gulf Stream area, $P$ dominates over $E$ by about 30%–50% in qualitative agreement with, for example, Trenberth et al. (2003). With increasing cyclone intensity, $E$ remains stable, while $P$ slightly grows by about 20% and amounts to $10^{-4}$ kg m$^{-2}$ s$^{-1}$ ($8.6$ mm day$^{-1}$) for the most intense cyclones. The water content change at this stage ranges from $25 \times 10^{-5}$ kg m$^{-2}$ s$^{-1}$ for the most intense cyclones to $28 \times 10^{-5}$ kg m$^{-2}$ s$^{-1}$ for the relatively weak ones. This implies the lateral advection of $30 \times 10^{-5}$ kg m$^{-2}$ s$^{-1}$ for all cyclone intensities. At $0.2 < \tau < 0.4$, both $E$ and $P$ decrease by about 5%–10% compared to the initial stage, while the temporal change of the water content weakens significantly (5–6 times compared to the initial stage) being $2 \times 10^{-5}$ and $5 \times 10^{-5}$ kg m$^{-2}$ s$^{-1}$ for the most intense and the weakest cyclones, respectively. This implies the lateral advection of about $0.5 \times 10^{-5}$ kg m$^{-2}$ s$^{-1}$ to $0.8 \times 10^{-5}$ kg m$^{-2}$ s$^{-1}$ for the intense and weak cyclones, respectively. At the stage of maximum cyclone development, $E$ decreases to $0.4 \times 10^{-5}$ kg m$^{-2}$ s$^{-1}$ (about $3.5$ mm day$^{-1}$) and $P$ dominates over $E$ by 30% for all intensities. The water content change at $0.4 < \tau < 0.6$ is slightly above zero and a small imbalance can be equally attributed to the advection and to the uncertainty. For cyclones located in the same region, $E$ reasonably depends on the wind speed and, thus, on the cyclone intensity $G$ (Fig. 9). Independence of $E$ from $G$ at the initial stage implied by Fig. 14a does not contradict Fig. 9, since here we used estimates of maximum intensity during the lifetime. These may not correlate necessarily with the magnitude of $G$ at the initial stage. Figure 14a also shows that $E$ does not depend on the cyclone intensity at $\tau = 0.4$–0.6 (when estimates of $G$ and $E$ are approximately collocated). However, the sample at $\tau = 0.4$–0.6 includes cyclones spread from the subtropics to the subpolar regions involving, for example, the LAB and SE cyclones (Figs. 1 and 12). This effect plays a dominant role in forming fluxes compared to cyclone intensity.

Figure 14b shows estimates of the water budget terms for the four cyclone types (Table 1, Fig. 12). For all cyclones at all stages $P$ dominates over $E$ with the $E - P$ differences being the largest for the LAB cyclones at $0.2 < \tau < 0.4$ and the smallest for the SE cyclones. During the initial stage the water content change is largest for the LAB cyclones (nearly $4.7 \times 10^{-4}$ kg m$^{-2}$ s$^{-1}$) and smallest for the NE cyclones. The large PW change at the initial stage is associated with the active advection of the water vapor into the cyclone domain. Strong differences between the balance estimates for the MID and NE cyclones during their initial stages can be partly explained by the longer migrations of the NE cyclones compared to the MID cyclones. At $\tau = 0.2$, many NE cyclones propagate as far as the mid-Atlantic, while most of the MID cyclones still remain in the Gulf Stream area. Similarly, we can explain the differences in the PW change between the LAB and SE cyclones at $\tau = 0–0.2$. Most of the LAB cyclones during this stage still remain near the Gulf Stream area where advection of the tropical dry air is quite strong, while the SE cyclones during the period $\tau = 0–0.2$ already pass distances twice as large compared to the LAB cyclones. At $0.2 < \tau < 0.4$ the smallest (close to zero) changes in PW are observed for the LAB cyclones with the largest changes being about $13 \times 10^{-5}$ kg m$^{-2}$ s$^{-1}$ (and the associated water vapor convergence of $16 \times 10^{-5}$ kg m$^{-2}$ s$^{-1}$) identified for the SE cyclones. This can be explained by strong advection of the tropical moisture in the low-level jets that may weaken the drying of these cyclones (Bao et al. 2006). At $0.4 < \tau < 0.6$, the water content change is slightly above zero for all cyclone types except for the LAB cyclones, for which it is slightly negative ($-3 \times 10^{-5}$ kg m$^{-2}$ s$^{-1}$). We have to note that this estimate is influenced by computational errors to a larger extent than those for the other cyclone types and should be taken with caution. For all cyclones at this stage the water budget is nearly balanced by $E - P$ and $\partial \text{PW}/\partial t$. The largest convergence of $3 \times 10^{-5}$ kg m$^{-2}$ s$^{-1}$ is identified for the SE cyclones. Summarizing, $E - P$ is negative for all cyclones and all stages, implying that cyclones always lose about 30% more water than they gain from the local evaporation. The strongest convergence is observed in the LAB cyclones at the initial stage and for the SE cyclones at $0.2 < \tau < 0.4$. During the cyclone’s decay stage, integral estimates (figure not shown)
are influenced by high uncertainties, which are associated with the potential overestimation of the size of the decaying cyclones. This may imply even physically unrealistic weak advection of water out of the cyclone domain, which makes it difficult to construct accurate estimates at the decay stage.

It is difficult to explicitly estimate the error term in (4), since the advection is not calculated directly. We can hypothesize that the errors result primarily from the inaccuracy of the estimation of the cyclone size. Uncertainties in determining the cyclone edge may result in the errors of both signs in all terms of (4). The methodology of Rudeva and Gulev (2007) has an accuracy of ±10% for oceanic cyclones. We performed sensitivity experiments, decreasing and enlarging the cyclone size by 5% and 10% by the corresponding proportional change of all radii. Then, the water budget terms (4) were recomputed for a new cyclone geometry. Table 5 shows that in most cases changes in the magnitude of different terms, associated with the increase–decrease of the cyclone size, are within the range of ±10% of the reference value, except for $\langle \partial PW/\partial t \rangle$ at $\tau = 0.4–0.6$ under $|\delta S_{\text{mod}} - S|/S = 10\%$, when the change in the magnitude of this term exceeds 20%. The smallest changes are identified for surface evaporation (4–10 times smaller compared to the other terms). At the earlier stage the larger changes are associated with the decrease in cyclone size, while at $\tau = 0.4–0.6$ the changes resulting from the increase in size become larger. We can conclude that the errors associated with the uncertainty of estimation of the cyclone size contribute up to 10% to the imbalance in

Fig. 14. Integrated components of the cyclone water vapor balance ($10^{-5}$ kg m$^{-2}$ s$^{-1}$) for (a) cyclones of different intensities and (b) for different cyclone classes at different stages during the life cycle: $0 < \tau < 0.2$, $0.4 < \tau < 0.6$, and $0.8 < \tau < 0.9$. In (a) red bars stand for $G < 20$ hPa, blue bars stand for $20 < G < 35$ hPa, and green bars stand for $G > 35$ hPa. In (b) the bars stand for LAB (red), MID (blue), NE (green), and SE (peach).
most cases, being up to 20% at \( \tau = 0.4–0.6 \). Thus, the magnitude of the error relative to the advection can be roughly estimated as 10%–20%.

8. The role of cyclones in forming integral air–sea fluxes over the North Atlantic

Our analysis implies that the strongly positive fluxes in the rear part of the cyclone are always associated with the weak evaporation and the negative sensible fluxes in the cyclone’s forward part. It is interesting to estimate the net role of cyclones in forming the surface fluxes integrated over the North Atlantic. We performed the estimates of the total turbulent heat fluxes associated with cyclones by integrating the fluxes over the cyclone area at every time step and over the lifetime. These estimates were compared with the total surface turbulent fluxes over the North Atlantic (30°–70°N) in order to answer the question whether cyclones really provide much higher than the average air–sea turbulent fluxes. Our estimates show that on average the winter cyclones per se (as they are defined here) do not provide much higher turbulent fluxes compared to those averaged over the North Atlantic. The mean sensible heat flux for the whole North Atlantic is 45 W m\(^{-2}\) and the mean \( Q_s \) produced by cyclones is 50 W m\(^{-2}\). Similarly, for the latent heat flux, we obtained the value of 115 W m\(^{-2}\) for the whole North Atlantic and 110 W m\(^{-2}\) for the area exposed to the cyclones during their lifetimes. However, Figs. 11 and 13 imply that at least at the initial stage cyclones do provide surface turbulent fluxes characterized by considerably higher values than the background fluxes. Figure 15 shows the scatterplot between the two ratios:

\[
\lambda_s = S_{\text{track}}/S_{\text{NA}}, \quad \lambda_q = Q_{\text{track}}/Q_{\text{NA}},
\]

where \( S_{\text{track}} \) is the area of the ocean exposed to the cyclone integrated over its lifetime, \( S_{\text{NA}} \) is the area of the North Atlantic (30°–70°N), \( Q_{\text{track}} \) is the flux integrated over the ocean area exposed to the cyclone, and \( Q_{\text{NA}} \) is the flux integrated over the North Atlantic. On average during their life cycle cyclones do not exhibit any excessive (compared to the ratio of areas) air–sea turbulent fluxes with about 40% of the points lying below the line \( \lambda_s = \lambda_q \). Our analysis of cyclone intensities shows that the fluxes above the average are primarily provided by the most intense cyclones. Figure 15b shows the histogram of the heat fluxes averaged over the North Atlantic area exposed to the propagating cyclones. About 43% of the cyclones show a surface flux larger than that averaged over the North Atlantic. Figure 15c shows the spatial distribution of the cyclone occurrence for which the integrated turbulent fluxes were larger than the mean North Atlantic fluxes. Clearly, most of these cyclones were identified over the Gulf Stream area with a few propagating in the central and eastern Atlantic. The occurrences in the Gulf Stream area are 3–7 times higher compared to those in the middle Atlantic. Thus, we can conclude that in the Gulf Stream

<table>
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<tr>
<th>( \tau = 0.2 )</th>
<th>( E )</th>
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<th>( \phi_{\text{PW/}t} )</th>
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<tbody>
<tr>
<td>90%</td>
<td>6.20</td>
<td>–9.88</td>
<td>24.55</td>
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<tr>
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<td>–9.23</td>
<td>26.33</td>
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<td>29.37</td>
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<td>–0.83</td>
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<tr>
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<td>6.11</td>
<td>–8.79</td>
<td>26.92</td>
<td>29.73</td>
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<td>6.97</td>
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<td>5.97</td>
<td>–7.68</td>
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<th>( \tau = 0.4–0.6 )</th>
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<tr>
<td>90%</td>
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area the dominant forcing of the turbulent fluxes is provided by synoptic systems. However, on average the North Atlantic midlatitudinal cyclones in winter do not provide any excessive (compared to the background values) turbulent heat fluxes from the ocean to the atmosphere.

9. Summary and discussion

We used an advanced cyclone compositing procedure for the analysis of winter oceanic cyclones in the North Atlantic from the NCEP–NCAR reanalysis dataset. The methodology is based on the cyclone transforming in a nondimensional azimuthal coordinate system, which allows for the effective comparison of different cyclones at various stages of cyclone development. The 576 cyclones selected were typical oceanic systems that were generated in the Gulf Stream area and then propagated over the North Atlantic. In the region of cyclogenesis maximum turbulent heat fluxes, the PW and precipitation in the cyclones clearly increase with the cyclone intensity quantified through the radial SLP difference. This is consistent with Bauer and Del Genio (2006), Chang and Song (2006), and Field and Wood (2007), who all agree that the cyclone intensity drives cyclone moisture characteristics. The evolution of the composites during the life cycle has shown a gradual cyclone drying trend with decreases in the PW and precipitation of about 40% and 50%–70%.

![Fig. 15. (a) Scatterplot of the ratios \( \lambda_q \) and \( \lambda_s \) for all selected cyclones. (b) Histogram of the heat fluxes averaged over the ocean area exposed to the propagating cyclones. (c) Distribution of the occurrences of cyclone centers when integrated over the cyclone area turbulent fluxes are larger than the North Atlantic mean fluxes.](image-url)
respectively, during their lifetime. The composites of the
turbulent heat fluxes imply the strongly positive \( Q_h \) and \( Q_e \)
in the rear part of the cyclones with close to zero fluxes in the
forward part and the decreasing gradient of the surface
fluxes between the rear and the forward sectors during the
life cycle. Although the composites of all characteristics
display high case-by-case variability, the differences be-
tween the cyclones of various intensities and between the
cyclones propagating to different regions are statistically
significant, implying the robustness of the composites.
Analysis of different cyclone types shows the strongest
turbulent fluxes for the MID and NE cyclones and the
smallest \( Q_{be} \) for the LAB cyclones. The strongest
changes in \( P \) during the lifetime were identified for the
LAB cyclones; the smallest change is shown by the SE
cyclones. Changes in precipitation are quite similar for
all cyclone types showing a strong decrease in \( P \) during
the lifetime.

Our results can be discussed from a number of view-
points. Since the composites were based on the NCEP–
NCAR reanalysis dataset, the findings are largely affected
by the problems inherent in this reanalysis system. Ex-
cessive (compared to the satellite observations) winter
precipitation over the midlatitudinal North Atlantic in the
NCEP–NCAR reanalysis was reported by Kistler et al. (2001),
Kanamitsu et al. (2002), and Bé ranger et al. (2006). It is generally attributed to the spinup effect [al-
though it is not as large as in the ERA reanalysis (Uppala
et al. 2005)] and to the parameterization of the horizontal
moisture diffusion. At the same time, the winter mid-
latitudinal turbulent latent heat flux in the North Atlantic
is also excessively high in the NCEP–NCAR reanalysis
(WGASF Group 2000; Moore and Renfrew 2002). This
bias can partly compensate for the excessive precipita-
tion, providing some realism in the representation of the
water cycle. Bauer and Del Genio (2006) analyzed precipi-
tation composites at the maximum cyclone development
stage in the northwest Atlantic using Tropical Rainfall
Measuring Mission (TRMM), ERA-40, and NCEP–
NCAR data and reported that both the ERA-40 and
NCEP–NCAR results underestimate precipitation, espe-
cially in the relatively weak cyclones. Many features of
the cyclone evolution revealed by the NCEP–NCAR re-
analysis may be different in the other products. However,
our methodology provides a good technique for the eval-
uation of these products.

We concluded that cyclones per se do not provide
considerably larger turbulent fluxes compared to the
background fluxes over the North Atlantic. This certainly
does not come as a surprise. Yuan et al. (2009) and Patoux
et al. (2009) obtained similar results for the Southern
Ocean using storm tracking based on satellite data and
surface fluxes computed with version 3 of the Coupled
Ocean–Atmosphere Response Experiment (COARE-3)
algorith (Fairall et al. 2003) from the ECMWF analyses
state variables. They reported that the ocean loses more
sensible and latent heat outside of cyclones. The effects of
cyclones in forming extreme air–sea fluxes can be more
likely connected with the circulation conditions outside of
cyclones. Extremely high turbulent fluxes can be observed
in the cyclone–anticyclone transition zones and the for-
ward parts of anticyclones following the propagating cy-
clones. From the viewpoint of air–sea interaction, these
conditions are not significantly different from those in the
cyclone cold sectors. Many case studies of cold-air out-
breaks (Chang et al. 1987; Bane and Osgood 1989; Yau and
Jean 1989; Fosdick and Smith 1991; Neiman et al. 1993;
Liu et al. 1997) showed extremely high fluxes exactly in the
cyclone–anticyclone transition zones. EOF-based compos-
it ing performed by Alexander and Scott (1997) and Zolina
and Gulev (2003) also showed that copropagating anom-
ali edes of turbulent fluxes are clearly associated with the
transition zones.

At the stage of cyclone deepening the moisture conver-
gence is provided by the advection from the areas outside
of the cyclone, which further transports the moisture along
the storm track. However, our analysis shows that cyclones
generated in the Gulf Stream area are unlikely to seriously
affect the European weather as they bring neither much
heat nor moisture. Most of these cyclones do not reach
Europe or, if they do, their effects upon it are weak. Also,
cyclones dry considerably during their propagation over
the North Atlantic. This process is likely to be associated
with the advection of the dry and cold air in the cyclone’s
rear part and with the rising of moisture accompanied by
the latent heat flux release in the warm conveyor belt. The
advective mechanism is more important at the initial stage
of the life cycle and for the cyclones moving northward. At
the same time, at the later stages, especially for the NE and
SE cyclones, the conveyor belt mechanism is likely to
dominate. The drying process is less pronounced for the
SE cyclones, which can involve the tropical moisture at the
low levels as described for the Pacific by Bao et al. (2006).
On approach to Europe, the drier cyclones can provide
only a weak or moderate level of precipitation, although
our estimates of all components of the water budget at the
latest stages of cyclone development are less robust com-
pared to the cyclone’s deepening stage. Recently, Dacre
and Gray (2009) argued that most of the cyclones that
cross western Europe originate in the eastern Atlantic
where the baroclinicity and the sea surface temperature
gradients are weak compared to the western Atlantic. In
this respect, cyclones that are generated in the Gulf Stream
area can provide some preconditioning for the generation
of the eastern Atlantic cyclones affecting the European
weather. This was also shown by Gulev et al. (2002), who
found a close association between the Gulf Stream SST gradients and the intensity of mesoscale and synoptic scale processes in the eastern Atlantic.

With regard to the further development of this study, we envision several possible lines. The advanced methodology of cyclone compositing provides a good basis for similar analyses of the alternative reanalysis products [e.g., ERA-40, ERA-Interim, the Japanese 25-yr Re-Analysis (JRA-25)], which would allow for intercomparison and cross validation of cyclone characteristics in different datasets. It is also attractive to extend the analysis to precipitation and surface flux fields obtained from in situ and satellite observations, which will help to validate the results based on NWP products. In this respect the study of Chang and Song (2006) [who used precipitation derived from International Comprehensive Ocean–Atmosphere Data Set (ICOADS) weather reports] provides a permissible prospect especially with the current availability of daily surface turbulent fluxes [e.g., the National Oceanography Centre climatology (NOC-2.0); Berry and Kent (2009)] and blended products (Yu and Weller 2007; Large and Yeager 2009), as well as of satellite precipitation provided by the Global Precipitation Climatology Project (GPCP) and Climate Prediction Center’s morphing technique (CMORF) at daily and higher temporal resolutions. It would be interesting to elaborate on the water cycle evolution in cyclones using a potential vorticity framework (Lackmann 2002; Brennan et al. 2008), which may help us to discover the importance of different mechanisms of moisture transport at various stages of the cyclone life cycle. Finally, the analysis of climate variability in cyclone composites could be of interest. Bengtsson et al. (2009), by employing a simplified compositing methodology, showed that the composites can significantly change in the warmer climate. Thus, it is very challenging to accurately quantify the observed changes in the cyclone characteristics using an advanced compositing methodology and multisource data.

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