Extreme upper level cyclonic vorticity events in relation to the Southern Hemisphere jet stream

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The mean seasonal variation in the frequency of occurrence of extreme upper level cyclonic vorticity events, and its relation to the jet stream, is examined in the Southern Hemisphere. During the austral summer to fall, extreme cyclonic vorticity occurs most frequently at the upper level jet stream core, while during the austral winter to spring, there is a main peak on the poleward flank of the subtropical jet, and a secondary peak on the poleward flank of the eddy driven jet. Composite analysis shows that the extremes in both seasons are associated with wave breaking and the formation of elongated vorticity tongues. In summer, extreme events occur when waves propagating on the eddy driven jet break nonlinearly while in winter, extreme events occur when waves on the eddy driven jet interact with waves on the subtropical jet. In both seasons, these extreme upper level vorticity events are associated with significant positive precipitation anomalies and a pattern of alternating positive and negative surface temperature anomalies.
1. Introduction

The anomalous Hemispheric-wide winter conditions of the Northern Hemisphere winter of 2013-14 have raised interest in what controls the distribution and frequency of extreme events, and how those might change with the large scale atmospheric circulation. As pointed out by Wallace et al. [2014], a preliminary examination of the flow during this winter suggests the extreme weather conditions arose, at least partly, from large undulations in the mid-latitude jet stream and the stratospheric polar vortex, with no clear proven relation to global warming or Arctic sea-ice melting. In fact, for most types of extreme events, we do not understand their relation to the large scale atmospheric circulation well enough to be able to deduce how their distribution might change even if we could predict the future large scale circulation.

Intuitively, we expect the distribution of extreme events to be linked to the large scale flow. At the simplest level, extreme values of a given field are related to basic statistical quantities of its distribution, like the mean, standard deviation, and skewness. Thus extremes in flow-based quantities like wind speed and vorticity are clearly related to the jet stream (mean flow) and storms (variance). We also expect the jet stream and its large scale undulations to indirectly affect the distribution of extremes of other fields like temperature and precipitation, by influencing the location and evolution of synoptic storms. Indeed, various studies have examined the relation of extremes to large scale patterns of variability. A partial list includes the relation of surface extremes to the latitude of the North Atlantic jet [Mahlstein et al., 2012], to large scale modes of variability in the Northern Hemisphere [Donat et al., 2010; Kenyon and Hegerl, 2008, 2010; Scaife
et al., 2008; Franzke, 2013] and Southern Hemisphere [England et al., 2006; Ummenhofer and England, 2007], and to other large scale flow features [e.g. Raible, 2007; Buehler et al., 2011].

We will take a different focus, and examine the influence of one of the central characteristics of the global circulation - the jet stream type - on extreme events. Meteorologists have long noticed two types of jet streams - subtropical jets which form due to meridional advection of angular momentum by the Hadley circulation [e.g. Schneider, 1977; Held and Hou, 1980], and polar front jets, also referred to as eddy-driven jets, which form due to the convergence of momentum by the synoptic storms [e.g. Panetta, 1993]. The interaction of these two types of jets with synoptic storms is very different, resulting in very different latitudinal locations, vertical structures and temporal variations of these jets [Lee and Kim, 2003; Eichelberger and Hartmann, 2007]. Thus, while the thermally driven subtropical jet is strongly baroclinic and its latitudinal position is relatively constant with time, the eddy driven jet has a strong surface westerly component, and it meanders strongly in its latitudinal position.

The Southern Hemisphere jet stream undergoes a sharp seasonal transition, from a dominantly thermally driven subtropical jet during austral winter, to an eddy driven mid latitude jet during austral summer [e.g. Nakamura et al., 2004]. In this study we will make use of this transition, and examine how the distribution of extreme upper level vorticity events changes with season. This will provide at least an upper bound on the degree to which the large scale flow, in particular the location and strength of the jet stream, affect extremes. We choose to examine extreme values of upper level cyclonic vorticity because
we expect it to be more strongly affected by the large scale circulation compared to the
more traditional fields used for studying extreme weather events (e.g. surface temperature
or precipitation), while at the same time to be indicative of extreme surface weather.
Several studies have linked extreme precipitation and surface temperature conditions with
upper level flow features like Rossby wave breaking, potential vorticity (PV) streamers,
and deep upper level troughs [e.g. Martius et al., 2006, 2013; Romero et al., 1999; Jacobitz,
1987; Krichak et al., 2007; Massacand et al., 1998; Sprenger et al., 2012]). In these
studies, the influence of upper level flow features on surface conditions involves vorticity
dynamics. Though wave breaking and the formation of PV streamers clearly influence
the vorticity distribution, the explicit relation to extreme vorticity values has not been
examined before. We start by examining how the climatological frequency of occurrence
of extreme negative vorticity events varies spatially and seasonally with respect to the
type of jet stream. Then we show that the formation of these extreme events is indeed
associated with wave breaking. Further, the differences in the evolution of these extreme
events is examined for the two solstice seasons, again taking the jet stream structure and
location into account.

2. Data and diagnostics

In this paper we present analysis using daily mean 300 hPa wind fields from ERA
Interim [Dee et al., 2011] for years 1979-2012. Calculations were also done using NCEP
re-analysis [Kalnay et al., 1996] for years 1959-2012, and only results which are found for
both reanalyses are presented here. Vorticity $\zeta$ is calculated from the daily horizontal
winds \( (u \) in the longitudinal direction \( \lambda \) and \( v \) in the latitudinal direction \( \varphi \)):

\[
\zeta = \frac{1}{a_e \cos \varphi} \frac{\partial v}{\partial \lambda} - \frac{1}{a_e \cos \varphi} \frac{\partial (u \cos \varphi)}{\partial \varphi}
\] (1)

where \( a_e \) is the Earth’s radius. We define an extreme cyclonic vorticity event when \( \zeta \) drops below the lowest 1% of vorticity values of all days and all grid points within the Southern Hemisphere. We then calculate the frequency of occurrence of extreme events as a function of spatial location or day of the year. We use composite analysis to obtain a characteristic evolution of the extreme events in different seasons, at the latitude of most frequent occurrence. For extreme vorticity events which last more than one day in a row, we only choose the day with most extreme value. We also isolate the events spatially and temporally by discounting extreme values which occur within 30 degrees longitude and within five days of the event peak. We then center the events around the peak longitude.

Statistical significance is calculated using a bootstrap method. This is done by choosing 1000 random combinations of NComp days (the number of events in our composite) and longitudes to get a distribution of randomly chosen vorticity field composites. Those grid points in the composite for which the anomaly is outside the 1000-member random distribution are considered statistically significant (e.g. 99.9% significance). The composites are done both for full fields and for the anomalies, defined as the deviation from the climatological seasonal cycle. The statistical significance, however, is determined purely from the anomaly fields. The climatological seasonal cycle is calculated by averaging the given field for each calendar day over all years, and applying a 21 day running mean for smoothing.
3. Results

Figure 1 shows the seasonal-latitudinal distribution of the frequency of occurrence of extreme negative (cyclonic) vorticity events while Fig. 2 shows the spatial distribution of extreme events during the two solstice seasons (Jan-Mar and Jun-Aug). The climatological seasonal cycle (Fig. 1) was created by counting the number of extreme events which occurred at each latitude and each calendar date (excluding Feb 29th) while the spatial distribution maps (Fig. 2) were calculated by counting the number of days with extreme values at each grid point. The seasonal maps were divided by the number of years times the number of longitude grid points, and the spatial maps by the total number of days, to get a probability of occurrence. Thus a value of 0.015 at a given point on the plots (which is the minimum shading value) means that on average at the point’s latitude and day of year (Fig. 1) or latitude and longitude (Fig. 2), events occur 1.5% of the time (more frequently than the defining extreme event threshold of 1%). Also shown on both plots are the climatological zonal mean winds at 300 hPa and at 925 – 750 hPa, representing the upper level jet stream, and the lower level eddy driven jet respectively. The figures reveal a strong relation between the distribution of extreme events and the jet stream. We see a clear seasonal cycle which roughly follows the upper level jet stream (Fig. 1). During summer and fall (Dec-Apr, Fig. 1, Fig. 2 top), the upper and lower level jets coincide, indicative of a strong barotropic eddy-driven jet [e.g. Son and Lee, 2005]. The jets are also roughly zonally symmetric, and the extreme cyclonic vorticity events occur most frequently at this jet core. We note that the extreme vorticity events are concentrated at the jet core, despite the fact that the vorticity of the mean flow peaks at the jet flanks.
The picture is very different during winter and early spring (Jun-Oct, Fig.1, Fig.2 bottom) when the jet assumes a spiral structure. Over the Indian and Pacific oceans it has characteristics of a thermally-driven subtropical jet, while elsewhere, it has characteristics of a mid-latitude eddy driven jet [Nakamura and Shimpo, 2004]. In the zonal mean, we see an upper level subtropical jet (at the edge of the Hadley cell, not shown), alongside a mid-latitude lower level eddy-driven jet. In clear relation to the jet structure, the distribution of extreme cyclonic vorticity values splits with a main peak on the poleward flank of the upper level subtropical jet (confined like the jet to the Indian and Pacific Oceans), and a secondary peak on the poleward flank of the lower level eddy driven jet. Unlike in summer, the negative vorticity of the mean flow contributes partly to the extreme negative vorticity values. In particular, extreme vorticity anomalies (deviations from a seasonal cycle) have a single peak at the core of the subtropical jet, with no significant peak in higher latitudes (not shown).

To further examine the differences between the extreme events during these two time periods, we composite various fields around the extreme events at the peak latitude of the distribution, for the different winter and summer jet types. Thus we composite around Jun-Aug extreme cyclonic vorticity events at 33°S, between 50°E−250°E (the peak along the subtropical jet), and around Jan-Mar events at 49.5°S at all longitudes (the peak along the eddy-driven jet). The composites consist of 1521 extreme events during Jan-Mar and 1287 events during Jun-Aug (composites of the strongest 250 events yield similar results). Fig 3 shows the composites of 300 hPa geopotential height anomalies for days −6, −3, and 0, overlain on the centered composite of 300 hPa zonal winds, averaged over days −6
Fig 4 shows the days $-3$ and $0$ composites of 300 hPa absolute vorticity, a quantity which nicely tracks the overturning of vorticity contours during Rossby wave breaking events [Barnes and Hartmann, 2012], along with composites of precipitation and surface temperature anomalies at day 0.

Looking at the geopotential height and absolute vorticity composites, we see different flow evolution between the two seasons. During Jan-Mar we see an upper level wave packet which starts developing around day $-6$, and propagates on the jet stream while growing in amplitude (days $-3$, $0$), until it deforms into an elongated tongue of low vorticity which deepens at the event center (day 0). The deepening of the low absolute vorticity center (day 0) seen also in geopotential height, indicates a contribution from stretching (supported by composites of potential temperature, not shown). This suggests the absolute vorticity tongue is part of a PV streamer which forms during these events.

During Jun-Aug the extreme cyclonic vorticity also develops when an existing wave packet strengthens and deforms to form a vorticity tongue, but in contrast to the summer case, the extreme negative vorticity anomaly on the thermally driven subtropical jet involves interaction with waves on the midlatitude eddy driven jet. This is seen clearly in the geopotential height field. On day $-6$ we see a midlatitude wavepacket on the midlatitude jet at upstream longitudes of $-180^\circ$ to $-30^\circ$ with a subtropical wave packet of higher zonal wavenumber subsequently developing (day $-3$) and merging with the midlatitude wavepacket. This merging is coincident with the formation of a southward-elongated high-vorticity tongue, evolving to produce the extremely negative vorticity values by day 0. A very similar upper level geopotential height evolution was observed by Marengo...
et al. [2002], in relation to extreme winter cold events in Brazil, suggesting the upper level
vorticity anomalies are relevant also to surface weather.

A surface influence is indeed found for both seasons, as can be seen from the day 0
composites of surface temperature and precipitation (Fig. 4, right column). We see for
both seasons, significant localized negative (0.5 – 1.5°C) and positive (0.5 – 1°C) tem-
perature anomalies associated with northward and southward temperature advection to
the north-west and east of the extreme cyclonic vorticity center, and a significant precip-
itation anomaly of about 6 – 8mm/day to the north-east of the vorticity center. These
surface anomalies are consistent with the dynamical evolution around PV streamers [as
summarized in Schlemmer et al., 2010]. We also see, a weaker but significant positive
temperature anomaly (0.5°C) in high latitudes, west of the event center. During Jan-
Mar the surface temperature anomalies form a zonally aligned wave packet, spanning the
latitudes between 30°S – 60°S with one of the positive anomaly centers extending to
high latitudes (to 70°S at 60° west of the event center), while during Jun-Aug the wave
packet assumes a slanted orientation (more eastward in the subtropics), reaching further
into the tropics with significant negative anomalies of surface temperature and 300 hPa
geopotential height anomalies at high latitudes (65°S – 75°S) to the east and west of the
composite center. These patterns are consistent with the Jan-Mar event involving waves
on the eddy driven jet, and the Jun-Aug event involving waves on the midlatitude eddy
driven jet and subtropical thermally driven jet.
4. Conclusion

The spatial distribution of extreme upper level negative vorticity events varies seasonally in a manner which is clearly related to the seasonal evolution of the zonal jet stream. The spatial relation to the jet position also varies seasonally. During summer, when the jet is eddy driven and in the midlatitudes, the extreme events occur most frequently at the jet center. During winter, on the other hand, the upper level jet is in the subtropics, and strongly thermally driven, while the lower level jet is in the midlatitudes, and the extreme events occur most frequently on the poleward flank of the upper level subtropical jet, with a secondary peak on the poleward flank of the midlatitude eddy driven jet. This difference in position relative to the jet comes along with differences in the characteristic temporal evolution of the extreme vorticity events (as seen from a composite analysis). While during summer, waves propagating on the jet break and form a negative vorticity anomaly at the jet core, during winter, the wave breaking involves an interaction between anomalies on the subtropical and midlatitude jets, so that the extreme vorticity values form on the poleward flank of the subtropical jet.

We note that in both solstice seasons, at the latitude of peak occurrence of extremes, an extreme event occurs somewhere at that latitude almost every other day. The composite analysis, which is based on all of these events essentially tells us where and when in the evolution of typical wave packets, extreme vorticity values occur. Given the very different structure and evolution of wave packets between the two different seasons, we get a different location of vorticity extremes relative the the jet.
The differences in temporal evolution make sense given that the different jet types arise from different dynamical balances between the zonal jets, the midlatitude eddies and the Hadley circulation. The subtropical thermally driven jet is less baroclinically unstable because the meridional gradient of the Coriolis force ($\beta$) is larger in the subtropics [Lachmy and Harnik, 2014]. Moreover, the subtropical jet peaks at the edge of the Hadley cell, where the eddy momentum flux convergence is zero [e.g. Son and Lee, 2005]. Thus, this type of jet can only be sustained in the presence of weak enough eddies [Lachmy and Harnik, 2014]. Correspondingly, cyclogenesis and baroclinic growth (e.g. as indicated by transient eddy heat fluxes) are strongest in mid and high latitudes, and in particular, during winter, they are weakest at the longitudes where the subtropical jet is strongest [e.g Simmonds and Keay, 2000; Nakamura et al., 2004; Lachmy and Harnik, 2014]. The peak in the occurrence of extreme vorticity values at the subtropical jet during winter, despite the weaker wave growth, seems to be enabled only through an interaction with the midlatitude jet waves, but more studies are needed to test whether this is indeed so.

The relevance of extreme upper level vorticity to extreme surface weather needs to be further examined. Our composite analysis (Fig. 4, right plots) suggests extreme upper level cyclonic vorticity values are associated with significant surface temperature and precipitation anomalies. While the composited temperature anomalies are not very large, a few other studies have shown that the upper level flow patterns are related to the formation of extreme negative surface temperature anomalies in the Southern Hemisphere. In particular, Sprenger et al. [2012] and Marengo et al. [2002] found cold events in Brazil to be associated with wave breaking and PV streamer formation, and with upper level...
geopotential height patterns very similar to Fig. 3, while Ashcroft et al. [2009] found extreme Australian cold events to be associated with a temporary merging of the subtropical and mid latitude jet streams.

The relation between upper level vorticity extremes and wave breaking needs further study. During Jan-Mar, the elongated vorticity tongues found in our composites hardly have any tilt in the latitude-longitude plane, while during Jun-Aug they have a very weak tendency toward anti-cyclonic wrapping. Thus it is hard to determine from the composites if cyclonic or anti-cyclonic wave breaking is involved. We note that Berrisford et al. [2007] find wave breaking of both kinds in the inter-jet region during austral winter.

Several studies have examined the seasonal distribution of wave breaking in the Southern Hemisphere in relation to the jet streams. Hitchman and Huesmann [2007] find a peak in the frequency of wave breaking just polewards of the subtropical jet during austral winter (their Fig. 3), slightly poleward of the region of maximum extreme vorticity events (Fig. 2b). During austral summer, however, they find the strongest mid-latitude wave breaking just east of southern Australia where the jet is weakest, while we find relatively less extreme negative vorticity events at those longitudes. They, however, relate wave breaking with blocking highs, and not vorticity lows. Ndarana and Waugh [2011] examined the Southern Hemisphere seasonal cycle of wave breaking and found a clear equatorwards shifting of the region of maximum wave breaking during austral winter, consistent with the jet becoming subtropical. The seasonal and spatial variations in the frequency of wave breaking, however, depend on the theta surface chosen, and are hard
to relate further to the distribution of extreme vorticity events without a more detailed study.

Finally, a preliminary analyses of the extremes of other variables also suggests strong dynamical relations to the large scale flow, in particular the type of the jet stream. By nature of the problem, detecting and predicting future changes in the distribution of extreme events involves a high degree of uncertainty, and requires much longer observational records than other dynamical studies. In the future, a better understanding of the dynamical relations to the large scale flow can hopefully be incorporated to improve statistical studies and prediction of extreme events.

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References


Franzke, C. (2013), Persistent regimes and extreme events of the north atlantic atmos-
pheric circulation, Phil. Trans. R. Soc. A, 371, 20110,471.

Held, I. M., and A. Y. Hou (1980), Nonlinear axially-symmetric circulations in a nearly

Hitchman, M. H., and A. S. Huesmann (2007), A seasonal climatology of rossby wave
breaking in the 3202000-k layer., J. Atmos. Sci., 64, 19221940.

Jacobeit, J. (1987), Variations of trough positions and precipitation patterns in the

Soc., 77(434-471).

Kenyon, J., and G. C. Hegerl (2008), Influence of modes of climate variability on global

Kenyon, J., and G. C. Hegerl (2010), Influence of modes of climate variability on global

Krichak, S. O., P. Alpert, and M. Dayan (2007), A southeastern mediterranean PV
streamer and its role in december 2001 case with torrential rains in israel, Nat. Hazards
Earth Syst. Sci., 7, 21–32.

Lachmy, O., and N. Harnik (2014), The transition to a subtropical jet regime and its

Lee, S., and H. K. Kim (2003), The dynamical relationship between subtropical and


Figure 1. A daily-latitude plot of the fractional number of days and longitudinal grid points with extremely low (below the 1st percentile) 300 hPa vorticity values on a given calendar day and latitude, counting over all years and longitudes (colors). Only values above 0.015 are shaded. Also shown in contours are the climatological zonal mean zonal winds at 300 hPa (black, thin contours are 15, 20 m/sec, thick contours are 25, 30 m/sec) and the 925 – 750 hPa mean (green, contour interval 3 m/sec, only positive values are shown). The seasonal cycle is calculated by averaging each calendar day over all years, smoothed with a 21 day running mean.
Figure 2. The seasonal mean distribution of the fraction of the days with extreme 300 hPa vorticity values at a given grid point (colors) counting over all seasons between 1979-2012 for the periods: a) Jan-Mar, b) Jun-Aug. Also shown in contours on both graphs are the corresponding climatological seasonal mean zonal winds at 300 hPa (black, thin contour is 15 m/sec, thick contours are 20 : 5 : 40 m/sec) and the 925 – 750 hPa mean (green, contour interval 5 m/sec only positive values are plotted).
Figure 3. The longitudinally-centered composites of 300 hPa geopotential height anomalies for the most extreme negative vorticity events occurring at latitude 49.5°S during Jan-Mar (top row) and at latitude 33°S during Jun-Aug. Shown are the composites for days −6, −3, and 0 (left, middle and right, respectively), plotted on top of the zonal wind composite averaged over days −6 : 6. The shading marks the significant regions for geopotential height anomalies. Significance levels (99.9%) are determined based on a bootstrap method with respect to randomly chosen and centered vorticity fields. The Jan-Mar and Jun-Aug composites are based on 1521 and 1287 events respectively. Geopotential height contour intervals are ±(15, 50 : 50 : 150), positive in red and negative in blue. Wind contours are 25, 30, 35 m/sec in black. The center of the composites is marked by a magenta dot.
Figure 4. The longitudinally-centered composites of different fields, for the Jan-Mar and Jun-Aug extreme vorticity composites of Fig. 3 (top and bottom rows respectively). Shown are absolute vorticity at days $-3$ and $0$ (left and middle plots) and the surface temperature anomaly alongside the total precipitation at day $0$ (right plots). The shading marks the significant regions for absolute vorticity and surface temperature, while for precipitation, only significant values are plotted. Absolute vorticity contours are $(-1.4 : 0.2 : -0.4) \times 10^{-4} \ sec^{-1}$ from blue to red. The surface temperature contour interval is $0.3^\circ C$ (with contours spanning values of $-1.5^\circ C$ to $0.9^\circ C$). Negative temperature anomalies are in blue while positive anomalies are in red. Precipitation contours are $3, 4, 6, 8 \ mm/day$ in black, only significant values are shown.