Diagnosing inter-model variability of 20th century Northern Hemisphere Jet Portrayal in 17 CMIP3 Global Climate Models

Sharon C. Jaffe\textsuperscript{1,2}*, David J. Lorenz\textsuperscript{2}, Daniel J. Vimont\textsuperscript{1,2}, and Jonathan E. Martin\textsuperscript{1}

1. Department of Atmospheric and Oceanic Sciences, University of Wisconsin-Madison, Madison, WI, USA.

2. Nelson Institute Center for Climatic Research, University of Wisconsin-Madison, Madison, WI, USA.

\textsuperscript{*Corresponding author address:} Sharon C. Jaffe, Department of Atmospheric and Oceanic Sciences, University of Wisconsin-Madison 1225 W. Dayton St., Madison, WI 53706.

E-mail: scjaffe@uwalumni.com
The present study focuses on diagnosing inter-model variability of non-zonally averaged jet stream portrayal in 17 global climate models (GCMs) from the CMIP3 dataset. Compared to the reanalysis, the ensemble mean 300 hPa Atlantic jet is too zonally extended and located too far equatorward in GCMs. The Pacific jet varies significantly between modeling groups, with large biases in the vicinity of the jet exit region that cancel in the ensemble mean. After seeking relationships between 20th century model wind biases and (1) the internal modes of jet variability or (2) tropical sea surface temperatures (SSTs), it is found that biases in upper-level winds are strongly related to an ENSO-like pattern in winter mean tropical Pacific SST biases.

The temporal variability of the upper-level zonal winds in the 20th century is found to be accurately modeled in nearly all 17 GCMs. Also, it is shown that Pacific model biases in the longitude of EOF 1 and 2 are strongly linked to the modeled longitude of the Pacific jet exit, indicating that the improved characterization of the mean state of the Pacific jet will positively impact the modeled variability.

This work suggests that improvements in model portrayal of the tropical Pacific mean state may significantly advance the portrayal of the mean state of the Pacific and Atlantic jets, which will consequently improve the modeled jet stream variability in the Pacific. To complement these findings, a second paper is in progress that examines 21st century GCM projections of the non-zonally averaged NH jet streams.
1. Background

A poleward shift of the jet streams under anthropogenic climate change has been theorized (Chen and Held 2007; Lorenz and DeWeaver 2007; Kidston et al. 2011), observed over the past 30 years (Thompson and Wallace 2000; Marshall 2003; Hu et al. 2007; Johanson and Fu 2009), and is broadly expected to continue into the future (Solomon 2007). Despite the multitude of studies acknowledging this poleward shift, jet stream winds still vary significantly between observational datasets and various model simulations. The present study contributes a detailed analysis of global climate model (GCM) portrayal of jet stream structure and variability in 20th century simulations as the precursor to a second study that will analyze 21st century GCM projections of jet stream portrayal.

Jet streams are closely related to storm frequency and intensity across the mid-latitudes and a small change in jet position or intensity significantly impacts the weather experienced by a large fraction of the world’s population. Also, it is important that GCMs correctly model the large-scale circulation (of which the jet stream is a primary feature) in order to gain confidence in other variables that may be controlled by the large-scale circulation such as precipitation over North America related to the Pacific storm track. A careful examination of previous work on this topic reveals that many studies infer jet position based upon the poleward extent of the Hadley Cell, the phase of the hemispheric annular mode, or the position of mid-latitude storm tracks.

The poleward boundaries of the Hadley Cell effectively represent the latitudinal extent of the tropical atmosphere and are coincident with the locations of the subtropical jets. Observational studies exploring a variety of reanalysis and OLR datasets show that a 2° - 4.5° latitude expansion of the Hadley Cell has occurred between 1979-2005 (Hu et al. 2007). While this time period may not be long enough to distinguish a long-term trend from decadal variability, this observed widening does not seem to be explained by internal atmospheric variability, which is less than 1.5° latitude in preindustrial GCM experiments (Johanson and Fu 2009). GCMs project even more poleward expansion of the Hadley Cell
in the future, averaging to a 2° latitude expansion by the end of the 21st century (Lu et al. 2007). This estimate is much smaller than what has already been observed, suggesting that either the subtropical jet will translate poleward under anthropogenic climate change more than estimates by GCMs suggest or that much of the observed ”trend” is a result of internal variability.

While the Hadley Cell is used as a proxy for the subtropical jet, the northern/southern annular mode (NAM/SAM) (also called the Arctic/Antarctic Oscillation) describes a north-south shift of mass between the mid-latitudes and the poles, indirectly describing a north-south shift of the polar (i.e. eddy-driven) jets (Thompson and Wallace 2000). In addition, NAM/SAM are the dominant modes of hemispheric climate variability at all levels, making the annular mode a convenient and useful proxy for describing polar jet stream translations. In agreement with other measures of jet stream position, the observed NAM and SAM have trended positive over the latter half of the 20th century, indicating the occurrence of a poleward shift of the polar jet in both hemispheres (Thompson et al. 2000; Marshall 2003). However, the magnitude of this trend is currently in question because the annular mode has become significantly less positive since 2000 (Overland and Wang 2005). Also, a recent study suggests that using the sea-level NAM/SAM, as is common practice, is ineffectual to describe jet shifts because it does not take into account the baroclinic structure of the anthropogenic climate change signal (Woollings 2008). Despite this uncertainty, future projections of the NAM/SAM are certainly important. In fact, inter-model variance of the NAM in climate projections is shown to be responsible for up to 40% of surface temperature and precipitation variance over Eurasia and North America in late 21st century projections (Karpechko 2010).

The more common proxy for polar jet stream position is the mean position of mid-latitude storm tracks, which are dynamically tied to polar jet stream position and intensity (Valdes and Hoskins 1989; Orlanski 1998; Chang et al. 2002). Storm tracks, which are predominantly located on the downstream and poleward side of the polar jets, are well replicated in reanalysis datasets using both feature-tracking algorithms and statistical metrics (Bengtsson et al.
2006; Ulbrich et al. 2008). However, interpreting future projections of storm track position is much more difficult. While some studies find that modeled storm tracks shift poleward by the end of the 21st century (Yin 2005), other studies suggest a poleward expansion and intensification of future storm tracks (Wu et al. 2010). In general, the projected poleward shift of southern hemisphere (SH) storm tracks is much clearer and more robust than the shift of storm tracks in the northern hemisphere (NH), which are fraught with model discrepancies (Bengtsson et al. 2006; Ulbrich et al. 2008).

The few studies that look directly at jet stream winds have also lacked consensus with regard to modeled future jet stream structure, especially in the Northern Hemisphere. While the World Climate Research Programme’s (WCRP’s) Coupled Model Intercomparison Project phase 3 (CMIP3) multi-model dataset (Meehl et al. 2007) ensemble mean shows a poleward shift and intensification of zonal mean zonal winds (Lorenz and DeWeaver 2007; Kushner et al. 2001), the inter-model spread is larger than the ensemble mean, reducing confidence in GCM projections (Kidston and Gerber 2010; Woollings and Blackburn 2012). Overall, low-level wind speeds (such as 850 mb or 10 m) are more consistent between modeling groups, perhaps indicating that the polar eddy-driven jet, which penetrates into the lower troposphere, unlike the subtropical jet, has a clearer response to anthropogenic climate change (Woollings and Blackburn 2012).

Because of the possible differences between subtropical and polar jet responses to anthropogenic climate change, studies that do not use a zonal mean perspective have found distinct results for different local jet stream structures. For instance, one recent observational study has shown that the NH Atlantic jet has shifted northward while the NH Pacific (East Asian) jet has not (Yaocun and Daqing 2011). Even on a regional scale, however, the variations among GCM portrayals of jet stream structure are significant. Intensification of upper-level wind has been shown to be consistent among GCMs, while the possible projected poleward shift of jet stream winds in the NH Atlantic and Pacific regions varies widely among modeling groups (Ihara and Kushnir 2009).
Adding complexity to the situation, all CMIP3 GCMs have been found to position the zonal mean jet too far equatorward in both hemispheres in the 20th century when compared to reanalysis data (Kidston and Gerber 2010; Woollings and Blackburn 2012). Because 21st century projections of jet structure are correlated with 20th century jet model biases (Kidston and Gerber 2010), the next step toward understanding future jet stream structure is a careful analysis of 20th century model biases, as presented in this paper. The goal of this study is to understand why there is a lack of model consensus of NH jet structure in 20th century simulations. A follow-up study will then discuss how to use this knowledge of 20th century simulations to better understand 21st century projections. While analyses of the zonal mean wind are a good starting point for an examination of the large-scale circulation, this study goes one step further to look at the upper-level winds separately for the Atlantic and Pacific basins without the use of zonal averaging. This type of analysis is valuable because of the complex jet dynamics associated with the asymmetric NH circulation. It also adds insight into the potential mechanisms that underlie a poleward shift of the jet stream, as will be discussed in much greater depth in the second part of this study.

The present paper is organized as follows. Section 2 outlines the reanalysis and GCM data used in this study. Results of a detailed comparison between GCM simulations and reanalysis are presented in Section 3. These results include the analysis of ensemble mean winter biases as well as an examination of inter-model variations and the portrayal of jet stream variability in GCMs. Conclusions are found in Section 4.

2. Data and Methods

In this study, observations are used to establish a climatology of NH jet streams based upon the 1980-1999 mean winter zonal winds. These observations come from the NCEP/NCAR Reanalysis 1 dataset (Kalnay et al. 1996). Seventeen GCMs are assessed in comparison with the observations to determine the accuracy of jet stream characterization in each model.
The 17 GCMs come from the World Climate Research Programme’s (WCRP’s) Coupled Model Intercomparison Project phase 3 (CMIP3) multi-model dataset for the climate of the 20th century experiment (20c3m) (Meehl et al. 2007). Table 1 lists the models included in this study. These particular models are chosen because they provide the daily-resolved data required for this study.

The present study employs daily 300 hPa zonal wind data and monthly sea surface temperature (SST) data for 20 boreal winter seasons. A complementary analysis using 700 hPa zonal wind data (not discussed) is found to be in close agreement with the results at 300 hPa. Daily wind data are smoothed using a 5 day running mean for the period encompassing November 1 through March 31st of each winter from November 1979 to March 1999 (with leap days removed). The 17 GCMs vary in resolution from 1.125° latitude x 1.125° longitude (Model 1 - INGV-SXG) to 4° latitude x 5° longitude (Model 17 - INM-CM3.0). In order to directly compare model and reanalysis data, each model is linearly interpolated to 2.5° latitude x 2.5° longitude resolution. Resolution differences between models are not found to be related to the accuracy of jet stream portrayal.

To create the mean winter zonal wind (SST), the smoothed (monthly) data are averaged over NDJFM and over all 20 years of the data period. The seasonal cycle of zonal wind is created by averaging each smoothed day (i.e. pentad) over all 20 boreal winter seasons. Smoothed daily wind data (with the seasonal cycle removed) are used to perform EOF/PC analysis.

3. Results and Discussion

a. Jet portrayal in NCEP/NCAR reanalysis

Both the mean state and variability of the upper-level winds are examined in order to gain a full understanding of the NH jet streams in reanalysis data, which is then used to assess GCM accuracy. The reanalysis winter mean 300 hPa zonal wind for 1980-99 is shown
in Figure 1, with wind speed maxima located in the Pacific and Atlantic basins (hereafter called the Pacific and Atlantic jets). The Pacific jet extends from East Asia across the Pacific basin and the Atlantic jet extends from the central continental United States to the west coast of Europe, tilting northeastward across the Atlantic basin.

The dominant modes of variability of the reanalysis are identified through empirical orthogonal function/principal component (EOF/PC) analysis of the smoothed daily 300 hPa zonal wind field with the seasonal cycle removed. EOF/PC analysis is performed on the reanalysis data over the North Atlantic (120°W - 20°E, 22.5°N - 80°N) and the North Pacific (100°E - 120°W, 22.5°N - 80°N) basins for winter (NDJFM) 1980-99. All EOFs/PCs shown in this study have been found to be well separated from higher-order EOFs/PCs as determined by the methodology of North et al. (1982). The two dominant modes of variability for each basin are shown in Fig. 2 as regressions of the 300 hPa zonal wind field (0° - 80°N) onto the first and second PCs of the zonal wind.

In the Pacific, the primary mode of variability explains 15.9% of the variance in the upper-level zonal wind, with the dominant variant structure located near the jet exit region (Fig. 2a), indicating a strengthening and weakening of zonal winds in this region. This mode represents an extension or retraction of the upper-level jet (Jaffe et al. 2011). The secondary mode of variability in the Pacific explains 11% of the variance in the upper-level zonal wind and looks quite different from the primary mode of variability, with a dipole of variant structures straddling the jet axis near the jet exit region (Fig. 2b). This pattern represents a northward/southward shift of the jet near the exit region.

In the Atlantic the first mode of variability (21% of the total variance) resembles a northward/southward shift of the eastern half of the jet (Fig. 2c) and the second mode of variability (18% of the total variance) characterizes a strengthening/weakening of the zonal wind speeds in the jet core especially in the eastern half of the jet (Fig. 2d). Because the structure of the Atlantic jet includes a southeast-northwest oriented tilt from southeastern North America toward Great Britain, an additional level of complexity is added to the in-
terpretation of these patterns of variability. To be certain of the correct interpretation, a
composite analysis is performed, averaging over the smoothed daily data that have PCs 1-2
greater than 1 standard deviation (or less than -1 standard deviation). This threshold in-
cludes the pentads with the largest magnitude variability of the upper-level winds. Results
are shown in Fig. 3 with the perturbation winds associated with each composite in the left
panel and the full winds in the right panel for each case. The composite analysis supports
the interpretation that the primary mode of variability of the Atlantic jet is a more like a
northward/southward shift and the secondary mode of variability is more like a strengthen-
ing/weakening of the jet exit region. The northward/southward shift of the jet can be seen
by comparing Fig. 3b and Fig. 3d and the extension/retraction of the jet can be seen by
comparing Fig. 3f and Fig. 3h The composite retraction of the Atlantic jet (Fig. 3h) also
strongly resembles a blocking pattern over the Atlantic basin.

Although the two leading modes of variability in the Pacific and Atlantic basin are
opposite one another, they can be interpreted similarly. To add meaning to the discussion
of these modes of variability they will be referred to as the “Shift EOF” (Pacific EOF 2
and Atlantic EOF 1) and “Extend/Rtract EOF” (Pacific EOF 1 and Atlantic EOF 2)
throughout the remainder of the paper. It has been suggested that the reversal of the
primary and secondary modes of variability between the Pacific and Atlantic jets results
from differences in the orientation of the subtropical and eddy-driven jets in the two regions
(Eichelberger and Hartmann 2007). This is likely related to the distinct nature of the jet
in each region, with the upper-level winds in the Pacific dominated by the influence of the
subtropical jet stream, and the upper-level winds in the Atlantic influenced by both the
polar and subtropical jet streams (Lorenz and Hartmann 2003; Eichelberger and Hartmann
2007). The difference between these two regions is important and would not be apparent in
a zonal mean analysis of the NH jets.
b. GCM bias of the mean winter jet

Model bias is defined to be the difference between the GCM and reanalysis 300 hPa zonal wind ($U_{bias} = U_{GCM} - U_{reanalysis}$). A positive model bias indicates that modeled zonal wind speeds are too high in a given location and negative model bias shows where modeled zonal wind speeds are too low. Figure 4a shows the average model bias for the 17 GCMs being considered. Overall, model bias of the upper-level zonal wind is on the same order of magnitude as the two dominant modes of variability seen in Fig. 2. Also, the amplitude of the average model bias is of the same order of magnitude as the standard deviation of model bias around the ensemble mean (Fig. 4b).

The largest bias of the model mean occurs in the Atlantic, where the jet is too extended and positioned too far equatorward on average in the models. The modeled jet is also positioned too far equatorward in the Southern Hemisphere (not shown), supporting the results of Kidston and Gerber (2010). The bias of the model mean in the Atlantic is somewhat larger than the standard deviation about the ensemble mean, indicating that biases are fairly consistent across models in this basin. The standard deviation for the Atlantic jet is positioned farther west than the mean model bias, with two maxima located on the poleward and equatorward flanks of the Atlantic jet.

Compared to the Atlantic, the bias of the model mean is small in the Pacific, with one isolated region of positive bias in the Eastern Pacific and another weak region of positive bias on the poleward flank of the Pacific jet. However, the standard deviation of models about the ensemble mean is quite large, indicating that models exhibit much variability in their portrayal of the Pacific jet. The standard deviation is largest, more than 8 m s$^{-1}$, in the Pacific jet exit region, strongly resembling the dominant mode of variability (EOF 1 - Extension/Retraction) for the Pacific region.

An examination of individual model portrayals of the mean winter jets (representative examples shown in Fig. 5) shows that some models have very small biases (Figs. 5a, b) while other models have bias patterns that resemble the dominant modes of variability of
As a first step toward understanding the cause of model biases, it is important to determine the relationship of these biases with the dominant modes of variability in the Pacific and Atlantic regions. Such relationships offer clues that help explain the existence of model biases. A normalized projection of each model’s mean winter bias onto the first and second EOFs of the upper-level zonal wind from the reanalysis (shown in Fig. 2) is used to quantify the relationship between the model biases and the observed modes of jet variability. This analysis is done separately for the Atlantic (120°W - 20°E) and Pacific (100°E - 120°W) basins and results are shown in Fig. 6. The sign convention used for the EOF 1-2 patterns of the reanalysis is that shown in Fig. 2. The position of each point with respect to the x-axis (y-axis) shows the value of each model’s projection onto EOF 1 (2) from the reanalysis.

In the Pacific, the projection of each model’s bias onto the observed dominant modes of variability shows two clusters of models. The first group of models (Group 1, depicted with crosses) have biases clustered on the negative x-axis, indicating their similarity to a retraction of the Pacific jet (negative EOF 1). The second group of models (Group 2, depicted with asterisks) have biases clustered on the positive x-axis, indicating their similarity to an extension of the Pacific jet (positive EOF 1). These two groups of models also have different positions with respect to the y-axis, with Group 1 being more likely to have a bias resembling a northward shift of the Pacific jet (positive EOF 2) and Group 2 being more likely to have a bias resembling a southward shift of the Pacific jet (negative EOF 2). It is important to note that despite the separation in EOF 1-2 space, the length of the normalized projections are only ∼0.5-6, indicating that EOF 1-2 are not complete in their explanation of jet bias. The black line connects the average Group 1 projection to the average Group 2 projection, showing the axis along which a combination of EOF 1 and 2 explain the inter-model bias of the upper-level zonal wind. The segregation of models along this axis is unusual and merits further examination. The open circles and diamonds in Fig. 6 will be explained in relation to other results discussed in section 3c.
In the Atlantic (Fig. 6b), models biases are more uniform, without the distinctive two
group structure found for the Pacific. Atlantic model biases mostly cluster in quadrant 4,
resembling both an extension and southward shift of the Atlantic jet (negative EOF 1 and
positive EOF 2). In Fig. 6b Atlantic models continue to be depicted as crosses/asterisks
according to their respective groups as determined for the Pacific – yet a delineation is
still apparent between Group 1 and Group 2. This delineation is shown by the fact that
crosses and asterisks barely overlap despite the fact that they are all located in the vicinity
of quadrant 4 in Fig. 6b. The fact that this grouped bias structure holds true in the Atlantic
suggests that despite the differences between the two basins, model biases in the Atlantic
and Pacific regions are likely linked.

In order to uncover the difference in spatial structure between Group 1 and Group 2, the
mean model bias of each group is calculated. The difference between Group 1 and Group
2 (Group 2 – Group 1) is shown in Fig. 7. There are two maxima, which indicate regions
of oppositely-signed bias between Group 1 and 2. The first (and largest) maximum in bias
difference is found in the Pacific jet exit region, in the same location as the large value of the
standard deviation of model bias shown in Fig. 4b. The second maximum in bias difference
is found on the southern flank of the Atlantic jet, also found in a location of high standard
deviation of model bias, as shown in Fig. 4b. No significant difference between Group 1 and
Group 2 is found in the Southern Hemisphere (not shown)

c. Relationship between jet bias and tropical Pacific SST bias in GCMs

The strong link between biases in the Pacific and Atlantic basins (despite different jet
dynamics in each region) suggests that a forcing external to the mid-latitude eddy/jet system
is involved in producing these model biases. Due to the far reaching influence of tropical
Pacific SST variations [e.g. the El Niño/Southern Oscillation (ENSO) phenomenon], this
region will be considered as a possible external forcing influencing model biases of mid-
latitude upper-level winds. This potential relationship will be examined using Maximum
Covariance Analysis (MCA).

MCA is used here to assess the dominant patterns of covariability between tropical SST biases and upper-level zonal wind biases in the same models. This technique identifies pairs of patterns that maximize the squared covariance between two variables: in this case the mid-latitude 300 hPa zonal wind (100°E - 20°W, 10°N - 80°N) and the tropical Pacific SST (30°S - 30°N, 120°E - 290°E). The covariance is identified across a given sampling dimension. Typically sampling is performed across time, but in this case sampling is done across the 17 GCMs to identify structures linked to model bias. Further explanation of MCA may be found in Bretherton et al. (1992), Wallace et al. (1992), and Deser and Timlin (1997). It is important to note that because this MCA analysis samples across model space instead of across time, ENSO-like SST patterns that are identified are not equivalent to inter-annual variability in any model. Instead, these ENSO-like patterns of SST show the winter mean state of the tropical Pacific that is associated with a given mode of inter-model covariability.

The first mode of covariability between the wind and SST explains 52% of the total squared covariance between the two fields. Considering that the second and third modes of covariability explain 16% and 13% of the total covariance respectively, the first mode is clearly dominant. Confirming the validity of this technique, the normalized root mean squared covariance (NRMSC) is calculated to be 0.30, meaning that there is a significant amount of total covariance between these two fields. In addition, the correlation between the two expansion coefficients (i.e. the left and right singular vectors) is 0.81, verifying that there is a high degree of coupling between the patterns identified in the wind and SST fields (Fig. 8c). Therefore, the first pattern of covariability identified by MCA appears robust and is shown in Fig. 8.

The patterns of covariability produced by MCA are depicted via regressing SST bias (homogeneous; Fig. 8b) and zonal wind bias (heterogeneous; Fig. 8a) onto the SST expansion coefficient. Regression onto the zonal wind expansion coefficient yields similar structures. Here we focus on the SST expansion coefficient as a potential predictor of zonal wind bias as
our leading hypothesis is that the tropical SST bias forces the zonal wind bias (further discussion found in Section 4). A scatter plot of the SST and zonal wind expansion coefficients is depicted in Fig. 8c, and demonstrates their strong correlation.

The homogeneous SST field (Fig. 8b) strongly resembles the positive phase of ENSO and exhibits a high spatial correlation with the observed ENSO SST pattern, shown in Fig. 9b ($r = 0.70$), with further discussion to follow. The heterogeneous wind field (Fig. 8a) is similarly spatially correlated with the grouped model bias shown in Fig. 7 ($r = 0.81$). It is notable that ENSO-like SST biases are so directly connected to mid-latitude jet biases through the first mode of covariability produced by MCA. This indicates that the portrayal of the winter mean state in the tropical Pacific affects the modeled upper-level midlatitude zonal winds in both the Pacific and Atlantic regions, suggesting that differences in NH jet stream portrayal between the 17 GCMs are primarily related to their respective representations of ENSO-like mean SST in the tropical Pacific. Again, we note that the causality is not established by the MCA; further discussion is found in Section 4.

The open circles added to Fig. 6, seek to explain the relationship between the Grouped model bias and the ENSO-like structure of tropical Pacific SST biases. The open circles show the values of the normalized projection of the heterogeneous wind pattern in the Pacific basin (Fig. 8a) onto the primary and secondary modes of zonal wind variability from the reanalysis data (EOF 1-2, Fig. 2a, b). Because of the non-signed nature of MCA, the open circles show both possible sign conventions. The addition of these circles show that the portion of the model wind bias due to ENSO-like mean SST biases in the tropical Pacific falls along almost the same axis as the bias of the jet stream between models in the Pacific. This reinforces the hypothesis that the uncertainties in mean winter Pacific jet stream portrayal are caused by each model’s treatment of winter mean tropical Pacific SST and suggests that if models produced a more consistent tropical Pacific winter mean SST distribution, Pacific jet model biases would be more consistent. The same relationship is not as clear in the Atlantic, though it is noteworthy that the portion of zonal wind bias that is related to ENSO-like mean state
biases (the circles in Fig. 6b) do generally align along the “Group 1, Group 2” axis. This suggests that tropical Pacific mean state biases may be responsible for some portion of the bias of the Atlantic jet as well.

In order to further confirm and detail the relationship of the Pacific modeled ENSO-like tropical mean state and mid-latitude zonal wind biases, we examine the spatial structure of zonal wind variations associated with temporal ENSO variations in the observed record, and compare these results with the results of the MCA above. One commonly-used metric for defining ENSO is the cold tongue index (CTI; Zhang et al. (1997)), defined by the sea surface temperature (SST) anomaly pattern over the eastern equatorial Pacific (6°S - 6°N, 180° - 90°W). Fig. 9 shows the regression of the reanalysis wintertime (NDJFM; annually resolved) zonal wind and wintertime SST fields onto the reanalysis wintertime CTI for 1950-2009. The regression therefore represents the observed patterns of wintertime SST and upper-level zonal wind associated with a positive ENSO event. Fig. 9b shows the canonical positive ENSO (El Niño) SST signal of warming in the eastern equatorial Pacific and Fig. 9a shows the wintertime zonal wind teleconnection pattern associated with positive ENSO SST anomalies. The positive phase of ENSO is associated with increased wind speeds within a subtropical band (15°-30°N) stretching from the dateline to approximately 90°W.

The normalized projection of the observed ENSO teleconnection pattern (Fig. 9a) onto the primary and secondary modes of variability from the reanalysis (EOF 1-2, Fig. 2a,b) is shown by the open diamonds added to Fig. 6. In the Pacific region (Fig. 6a), these diamonds also fall along nearly the same axis as inter-model variations in jet stream biases and the heterogeneous wind pattern produced by MCA. The near-alignment of these different variables shows that they all project onto a similar combination of EOF 1-2. Thus, it is even more likely that the ENSO-like bias of modeled SSTs explains inter-model differences in the bias of NH jet streams in the Pacific. While the link between Atlantic jet biases and ENSO is weaker, inter-model variations do lie along the same axis as the ENSO teleconnection pattern. Therefore, it seems that model biases in the portrayal of the Atlantic jet are also
affected by tropical Pacific mean state biases.

To further confirm the results of MCA, the mean winter upper-level wind field is regressed onto the mean winter CTI for each model, with results shown in Fig. 10. The results of this regression analysis look remarkably similar to the results of MCA, and are correlated with the heterogeneous wind field (Fig. 8a) at $r = 0.98$ and with the grouped model bias pattern (Fig. 7) at $r = 0.71$. This confirms that jet stream biases across the 17 GCMs are related to the ENSO-like biases in tropical Pacific SST in these models. However, a comparison between Fig. 10 and Fig. 9a shows that the jet stream bias pattern associated with ENSO-like SST biases in GCMs does not completely resemble the observed ENSO teleconnection pattern ($r = 0.46$). Additional thoughts on this issue are found in Section 4.

To quantify how much inter-model variance is explained by the ENSO-like pattern identified by MCA, the SST expansion coefficient (i.e. the left singular vector) of the first mode of MCA covariability is used as a predictor of inter-model variance of upper-level winds, allowing the determination of what percentage of the inter-model variance of upper-level winds is explained by GCM SST biases. The result of this analysis, shown in Fig. 11, finds that ENSO-like SST biases explain 21% of the NH inter-model variance of mid-latitude jet stream portrayal on average, with significantly more variance explained in the central subtropical Pacific and eastern subtropical Atlantic. Fig. 11b shows that while SST biases do not explain all of the inter-model variance in upper-level winds, they do explain a substantial portion, especially in the Pacific.

d. GCM portrayal of jet variability

To complete this analysis of NH jet stream portrayal, temporal variability of the upper-level winds is also considered. EOF/PC analysis is used to determine the primary and secondary modes of variability associated with the Pacific and Atlantic jets for each model. The same methodology is used as for the reanalysis data (Section 3a). EOF/PC 1 and 2 are well separated for all models.
Figure 12 shows the normalized projection of the first two EOFs of each model onto the first two EOFs of the reanalysis data (a. Pacific, b. Atlantic). Asterisks (Crosses) indicate the value of the projection of EOF 1 (EOF 2) of a given model onto EOF 1 and 2 of the reanalysis. For instance, an asterisk located at (1,0) would describe an exemplary model’s depiction of EOF 1 that is completely explained by reanalysis EOF 1 and not explained by reanalysis EOF 2. It is important to note that the sign of a given mode is arbitrary and therefore the polarity is assigned based upon the convention established by reanalysis EOF 1-2 (Fig. 2).

Most points cluster near (0,1) or (1,0), indicating that GCMs are successfully replicating the two dominant modes of variability. There are only 3 outlier points: two for the Atlantic and one for the Pacific. The Atlantic outliers are EOF 1 and 2 from Model 1 (INGV-SXG), and indicate the reversal of EOF 1 and 2 in that model (not shown). Because EOF 1 and 2 explain 15% and 14% of the variability of the Atlantic upper level winds respectively, this reversal is not a serious flaw in the modeled variability. For the one outlier in the Pacific (EOF 2 from Model 16, GISS-ER) it is found that EOF 2 and EOF 3 are reversed (not shown), another minor flaw since these modes explain 11% and 9% of the variability respectively. Therefore, all 17 models do a good job of replicating the two dominant modes of variability. Even the outliers have the correct structures represented in the wrong order. In fact, this variability appears more consistently replicated than the mean state of the jets in GCMs (Fig. 6).

Because the location of the perturbation wind speeds associated with the dominant modes of variability is located nearby the jet exit region (see, e.g., Fig. 2), a measure of the longitude of the jet exit region and longitude of wind speed anomalies associated with EOF 1-2 is used to find a functional relationship between jet mean state and variability.

A regression analysis, shown in Fig. 13 examines the relationship between the modeled modes of variability and modeled mean state of the Atlantic and Pacific jets. The longitude of the maximum wind perturbation associated with EOF 1-2 is regressed onto the longitude
of the jet exit region for each model, as defined by the local minimum of the zonal gradient of the mean winter zonal wind from each model. Model 16 (GISS-ER) is removed from the analysis of Pacific EOF 2 and Model 1 (INGV-SXG) is removed from the analysis of Atlantic EOF 1-2 because they do not correctly represent their respective EOFs (as shown in Fig. 12).

For the Pacific jet, the longitude of the maximum value of EOF 1-2 is highly correlated with the longitude of the jet exit region ($r = 0.88, r = 0.68$), shown in Fig. 13a, b (open circles show results from reanalysis dataset). The $y = x$ line is also shown for the Pacific, which indicates the line the regression would follow if EOF 1-2 were located exactly at the longitude of the jet exit region. Most models correctly portray the maximum wind perturbation to be located immediately downstream of the jet exit region.

Figures 13c, d show the regression analysis for the Atlantic jet, which does not display a similarly strong connection ($r = 0.41, r = 0.43$). The models all portray the maximum wind perturbation too far downstream for EOF 1 and most models portray the maximum wind perturbation too far upstream for EOF 2. Overall, there does not seem to be a link between the longitude of the mean state and variability for the Atlantic region, possibly because of the added complexity due to the southeast-northwest tilt of the jet in this region.

There is a robust correspondence between the Pacific jet mean state and its variability, but not the Atlantic jet mean state and variability. When these results are repeated using GCM 21st century A1B projections (not shown), this correspondence (or lack thereof in the Atlantic) does not change. Therefore, the Pacific jet exit region is critically related to the longitudinal position of EOF/PC 1 and 2, signifying that a correct characterization of the mean state of the Pacific jet stream is vital to producing an accurate portrayal of the variability of that jet stream.
4. Conclusions

This study has focused on determining the reliability and robustness of non-zonally averaged NH jet stream portrayal in 17 GCMs from the CMIP3 dataset. This work is motivated by previous studies showing that GCM projections of 21st century jet stream winds are related to biases in 20th century jet stream portrayal (Kidston and Gerber 2010). The results presented in this paper encourage targeted improvement of GCM jet stream portrayal, which is an important step toward assessing climate change impacts at a variety of scales.

Examination of mean model biases of upper-level zonal winds suggests that the modeled Atlantic jet is too zonally extended and located too far equatorward compared to the re-analysis. The ensemble mean Pacific jet has less bias than the Atlantic jet, but only because model agreement is much lower and biases in individual models cancel in the ensemble mean. Mean winter biases in both basins are significant compared to the observed variability of the upper-level zonal winds.

MCA and regression analysis are used in tandem to show that the NH biases in upper-level winds are strongly related to an ENSO-like pattern in winter mean tropical Pacific SSTs. Throughout the analysis we have implicitly assumed that tropical SSTs are responsible for forcing mid-latitude winds, suggesting that the variation in models’ portrayal of the tropical Pacific mean state contributes to the bias of the mid-latitude large-scale circulation. However, it is important to note that the reverse scenario is also possible. Recent studies have shown that variations in mid-latitude and subtropical winds may also conspire to produce tropical Pacific ENSO variations, as evidenced by the seasonal footprinting mechanism examined in Vimont et al. (2001). It is possible that this causal mechanism (from mid-latitude to the tropics) would also work in linking mid-latitude wind biases to biases in tropical Pacific SST. While the present study does not resolve that causality, the similarity between the spatial structure of ENSO’s teleconnection in the observed record to the model bias (Fig. 6a) suggests that biases in the tropical Pacific are influencing mid-latitude zonal wind biases. This proposed causality is also supported by recent findings that ocean circu-
lation uncertainties force uncertainties in the North Atlantic storm track in climate model
simulations (Woollings et al. 2012).

Temporal variability of the upper-level zonal winds is accurately modeled in nearly all
17 GCMs. Furthermore, it is shown that in the Pacific, biases that do exist in models’
portrayal of EOFs 1 and 2 are strongly linked to the modeled longitude of the jet exit in the
Pacific region. This result is particularly encouraging because it implies that an improved
characterization of the mean state of the Pacific jet will also positively impact the modeled
variability.

In conclusion, results herein indicate that improvements in model portrayal of the tropical
Pacific mean state may significantly advance the portrayal of the mean state of the Pacific
and Atlantic jets, which will consequently improve the modeled jet stream variability in
the Pacific. To complement these findings, a second paper examines 21st century GCM
projections of the non-zonally averaged NH jet streams. Those results show that ENSO-like
changes in the tropical Pacific mean state dominate inter-model variations in projections of
21st century NH jet streams.

Acknowledgments.

This research was supported by NSF grant ATM-0653795, NOAA grant NA08OAR4310880,
and NSF grant ATM-0806430. NCEP Reanalysis data was provided by the NOAA/OAR/ESRL
PSD, Boulder, Colorado, USA, from their Web site at http://www.esrl.noaa.gov/psd/. In
addition, we acknowledge the modeling groups, the Program for Climate Model Diagnosis
and Intercomparison (PCMDI), and the WCRP’s Working Group on Coupled Modelling
(WGCM) for their roles in making available the WCRP CMIP3 multi-model dataset. Sup-
port of this dataset is provided by the Office of Science, U.S. Department of Energy.
REFERENCES


Gordon, H. et al., 2002: The CSIRO Mk3 Climate System Model. Tech. rep., CSIRO.

Gualdi, S., E. Scoccimarro, A. Bellucci, A. Grezio, E. Manzini, and A. Navarra, 2006: The main features of the 20th century climate as simulated with the SXG coupled GCM. *Claris Newsletter*, **none** (4).


Kidston, J., G. Vallis, S. Dean, and J. Renwick, 2011: Can the increase in the eddy length scale under global warming cause the poleward shift of the jet streams? *J. Climate*, **24** (14), 3764–3780.


<table>
<thead>
<tr>
<th>Model</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>INGV-SXG</td>
<td>Gualdi et al. (2006) and Gualdi et al. (2008)</td>
</tr>
<tr>
<td>CSIRO-Mk3.0</td>
<td>Gordon et al. (2002)</td>
</tr>
<tr>
<td>CSIRO-Mk3.5</td>
<td>Gordon et al. (2002)</td>
</tr>
<tr>
<td>ECHAM5/MPI-OM</td>
<td>Jungclaus et al. (2006)</td>
</tr>
<tr>
<td>GFDL-CM2.0</td>
<td>Delworth et al. (2006) and Gnanadesikan et al. (2006)</td>
</tr>
<tr>
<td>BCCR-BCM2.0</td>
<td><a href="http://www.bjerknes.uib.no/">http://www.bjerknes.uib.no/</a></td>
</tr>
<tr>
<td>CGCM3.1 (T63)</td>
<td>Flato et al. (2000) and <a href="http://www.ec.gc.ca/ccmac-cccma">http://www.ec.gc.ca/ccmac-cccma</a></td>
</tr>
<tr>
<td>MRI-CGCM2.3.2</td>
<td>Yukimoto et al. (2006)</td>
</tr>
<tr>
<td>FGOALS-g1.0</td>
<td>Yu et al. (2002) and Yu et al. (2004)</td>
</tr>
<tr>
<td>CGCM3.1 (T47)</td>
<td>Flato et al. (2000) and <a href="http://www.ec.gc.ca/ccmac-cccma">http://www.ec.gc.ca/ccmac-cccma</a></td>
</tr>
<tr>
<td>ECHO-G</td>
<td><a href="http://mad.zmaw.de/Models/Modelliste1">http://mad.zmaw.de/Models/Modelliste1</a> neu.html</td>
</tr>
<tr>
<td>GISS-AOM</td>
<td>Russell et al. (1995) and Lucarini and Russell (2002)</td>
</tr>
<tr>
<td>GISS-ER</td>
<td>Schmidt et al. (2006)</td>
</tr>
<tr>
<td>INM-CM3.0</td>
<td>Volodin and Diansky (2004) and Galin et al. (2003)</td>
</tr>
</tbody>
</table>
List of Figures

1 Reanalysis zonal wind (m s\(^{-1}\)) at 300 hPa for NH winter (November-March) 1979–1999, contoured every 10 m s\(^{-1}\).

2 EOFs of the 300 hPa midlatitude zonal wind field (20°–80°N) regressed onto the total 300 hPa zonal wind field (0°–80°N). Solid (dashed) black lines represent positive (negative) perturbation isotachs, contoured every 4 m s\(^{-1}\) with the zero line removed for (a) Pacific basin EOF 1 (Extend/Retract), (b) Pacific basin EOF 2 (Shift), (c) Atlantic basin EOF 1 (Shift), (d) Atlantic basin EOF 2 (Extend/Retract). Gray contours show the 20 (30) m s\(^{-1}\) isotach of the mean 300 hPa zonal wind for the Atlantic (Pacific) basins, as in Fig. 1.

3 Composites of maximum variability of the 300 hPa zonal wind field in the Atlantic. Gray contours show the 20 m s\(^{-1}\) isotach of the mean 300 hPa zonal wind for the Atlantic. Solid (dashed) black lines in the left column indicate positive (negative) perturbation isotachs, contoured every 5 m s\(^{1}\) with the zero line removed. The right column shows perturbation isotachs added to the climatology in units of m s\(^{-1}\) for (a)-(b) 1st principal component (PC) greater 1 standard deviation (1*\(\sigma\)), (c)-(d) 1st PC less than -1*\(\sigma\), (e)-(f) 2nd PC greater than 1*\(\sigma\), (g)-(h) 2nd PC less than -1*\(\sigma\).

4 (a) Ensemble mean model bias and (b) Standard deviation of models about the ensemble mean for the 17 GCMs under consideration [m s\(^{-1}\)]. Solid (dashed) black contours in (a) represent positive (negative) values of ensemble mean bias, contoured every 1 m s\(^{-1}\) with the zero line removed. The gray contours show the 20, 40 m s\(^{-1}\) isotachs of the model mean winter 300 hPa zonal wind.
Solid (dashed) black contours show the positive (negative) bias of the 300 hPa zonal wind, contoured every 4 m s$^{-1}$ with the zero line removed for (a) Model 5: ECHAM5/MPI-OM, (b) Model 8: CGCM3.1 (T63), (c) Model 7: BCCR-BCM2.0, (d) Model 11: MRI-CGCM2.3.2, (e) Model 10: MIROC3.2 (medres), (f) Model 9: CNRM-CM3. Gray contours show the 20 (30) m s$^{-1}$ isotachs of the 300 hPa zonal wind for the Atlantic (Pacific) basin.

Normalized projection of the model bias of the 300 hPa zonal wind onto EOF 1 and 2 from the reanalysis for the (a) Pacific and (b) Atlantic basins. Models designated by crosses (+) are part of Group 1 and models designated by asterisks (*) are part of Group 2. Dashed circles indicate lines of constant correlation at $r = 0.25$, $0.5$, and $0.75$. The black line connects the average Group 1 projection to the average Group 2 projection. Open circles (diamonds) show the values of the normalized projection of the heterogeneous wind pattern shown in Fig. 8a (ENSO teleconnection pattern shown in Fig. 9a) onto EOF 1 and 2.

Solid (dashed) black lines show the positive (negative) model bias of the 300 hPa zonal wind, contoured every 1 m s$^{-1}$ with the zero line removed for (Group 2 – Group 1), where models are separated into Group 1 (models 1, 5, 7, 8, 9, 10, 12, 13, 17) and Group 2 (models 2, 3, 4, 6, 11, 14, 15, 16), based upon their delineation in Figure 6, as described in the text. Gray contours show the 20, 40 m s$^{-1}$ isotachs of the model mean 300 hPa zonal wind.

Results of MCA of tropical Pacific SSTs and mid-latitude 300 hPa zonal wind. (a) Heterogeneous wind regression map, (b) Homogeneous SST regression map, (c) Scatter plot of the wind and SST expansion coefficients. Solid (dashed) black contours in (a) represent positive (negative) perturbation isotachs, contoured every 1 m s$^{-1}$ with the zero line removed. Gray contours in (a) show the 20, 30 m s$^{-1}$ isotachs of the model mean 300 hPa zonal wind.
9 ENSO mid-latitude wind teleconnection (1950-2009) shown by: (a) Observed mean winter 300 hPa zonal wind regressed onto the mean winter CTI and (b) mean winter tropical Pacific SST regressed onto the mean winter CTI. Solid (dashed) black contours in (a) represent positive (negative) perturbation isotachs, contoured every 0.5 m s\(^{-1}\) with the zero line removed. Gray contours represent the 20, 30 m s\(^{-1}\) isotachs of the 300 hPa mean zonal wind.

10 Modeled mean winter cold tongue index (CTI) for each model regressed onto the modeled mean winter 300 hPa zonal wind field. Solid (dashed) black lines indicate positive (negative) perturbation isotachs, contoured every 1 m s\(^{-1}\) with the zero line removed. Gray contours show the 20, 30 m s\(^{-1}\) isotachs of the model mean 300 hPa zonal wind.

11 (a) Total inter-model variance of the mean winter 300 hPa zonal wind and (b) Inter-model variance of the mean winter 300 hPa zonal wind explained by the SST expansion coefficient of the first mode of MCA covariability. Variance is expressed in units of m\(^2\) s\(^{-2}\) and contoured every 10 m\(^2\) s\(^{-2}\)

12 Scatter plot of the normalized projection of each model’s PC/EOF 1-2 onto the corresponding reanalysis PC/EOF 1-2 for (a) the Pacific, (b) the Atlantic. EOF 1 is indicated by asterisks (*) and EOF 2 is indicated by crosses (+). Dashed circles indicate lines of constant correlation at r = 0.25, 0.5, and 0.75.

13 Longitude of the maximum wind perturbation associated with EOF 1/2 regressed onto the longitude of the jet exit region in each model for (a) Pacific EOF 1 (Extend/Retract), (b) Pacific EOF 2 (Shift), (c) Atlantic EOF 1 (Shift), (d) Atlantic EOF 2 (Extend/Retract). Open circles represent reanalysis data and the dashed line shows y=x. Longitudes are displayed in degrees east (e.g. 240°E = 120°W).
Fig. 1. Reanalysis zonal wind (m s\(^{-1}\)) at 300 hPa for NH winter (November-March) 1979–1999, contoured every 10 m s\(^{-1}\).
FIG. 2. EOFs of the 300 hPa midlatitude zonal wind field (20°–80°N) regressed onto the total 300 hPa zonal wind field (0°–80°N). Solid (dashed) black lines represent positive (negative) perturbation isotachs, contoured every 4 m s$^{-1}$ with the zero line removed for (a) Pacific basin EOF 1 (Extend/Retract), (b) Pacific basin EOF 2 (Shift), (c) Atlantic basin EOF 1 (Shift), (d) Atlantic basin EOF 2 (Extend/Retract). Gray contours show the 20 (30) m s$^{-1}$ isotach of the mean 300 hPa zonal wind for the Atlantic (Pacific) basins, as in Fig. 1.
Anomaly
a. (+) EOF 1 (Shift)
c. (ï) EOF 1 (Shift)
e. (+) EOF 2 (Extend/Retract)
g. (ï) EOF 2 (Extend/Retract)
10
10 ... (+) EOF 2 (Extend/Retract)
10
10 10
10
30
30
30
h. (ï) EOF 2 (Extend/Retract)

Fig. 3. Composites of maximum variability of the 300 hPa zonal wind field in the Atlantic. Gray contours show the 20 m s$^{-1}$ isotach of the mean 300 hPa zonal wind for the Atlantic. Solid (dashed) black lines in the left column indicate positive (negative) perturbation isotachs, contoured every 5 m s$^{-1}$ with the zero line removed. The right column shows perturbation isotachs added to the climatology in units of m s$^{-1}$ for (a)-(b) 1st principal component (PC) greater 1 standard deviation (1$\sigma$), (c)-(d) 1st PC less than -1$\sigma$, (e)-(f) 2nd PC greater than 1$\sigma$, (g)-(h) 2nd PC less than -1$\sigma$. 
Fig. 4. (a) Ensemble mean model bias and (b) Standard deviation of models about the ensemble mean for the 17 GCMs under consideration [m s$^{-1}$]. Solid (dashed) black contours in (a) represent positive (negative) values of ensemble mean bias, contoured every 1 m s$^{-1}$ with the zero line removed. The gray contours show the 20, 40 m s$^{-1}$ isotachs of the model mean winter 300 hPa zonal wind.
Fig. 5. Solid (dashed) black contours show the positive (negative) bias of the 300 hPa zonal wind, contoured every 4 m s\(^{-1}\) with the zero line removed for (a) Model 5: ECHAM5/MPI-OM, (b) Model 8: CGCM3.1 (T63), (c) Model 7: BCCR-BCM2.0, (d) Model 11: MRI-CGCM2.3.2, (e) Model 10: MIROC3.2 (medres), (f) Model 9: CNRM-CM3. Gray contours show the 20 (30) m s\(^{-1}\) isotachs of the 300 hPa zonal wind for the Atlantic (Pacific) basin.
Fig. 6. Normalized projection of the model bias of the 300 hPa zonal wind onto EOF 1 and 2 from the reanalysis for the (a) Pacific and (b) Atlantic basins. Models designated by crosses (+) are part of Group 1 and models designated by asterisks (*) are part of Group 2. Dashed circles indicate lines of constant correlation at \( r = 0.25, 0.5, \) and 0.75. The black line connects the average Group 1 projection to the average Group 2 projection. Open circles (diamonds) show the values of the normalized projection of the heterogeneous wind pattern shown in Fig. 8a (ENSO teleconnection pattern shown in Fig. 9a) onto EOF 1 and 2.
Fig. 7. Solid (dashed) black lines show the positive (negative) model bias of the 300 hPa zonal wind, contoured every 1 m s$^{-1}$ with the zero line removed for ($\text{Group } 2 - \text{Group } 1$), where models are separated into Group 1 (models 1, 5, 7, 8, 9, 10, 12, 13, 17) and Group 2 (models 2, 3, 4, 6, 11, 14, 15, 16), based upon their delineation in Figure 6, as described in the text. Gray contours show the 20, 40 m s$^{-1}$ isotachs of the model mean 300 hPa zonal wind.
Fig. 8. Results of MCA of tropical Pacific SSTs and mid-latitude 300 hPa zonal wind. (a) Heterogeneous wind regression map, (b) Homogeneous SST regression map, (c) Scatter plot of the wind and SST expansion coefficients. Solid (dashed) black contours in (a) represent positive (negative) perturbation isotachs, contoured every 1 m s$^{-1}$ with the zero line removed. Gray contours in (a) show the 20, 30 m s$^{-1}$ isotachs of the model mean 300 hPa zonal wind.
Fig. 9. ENSO mid-latitude wind teleconnection (1950–2009) shown by: (a) Observed mean winter 300 hPa zonal wind regressed onto the mean winter CTI and (b) mean winter tropical Pacific SST regressed onto the mean winter CTI. Solid (dashed) black contours in (a) represent positive (negative) perturbation isotachs, contoured every 0.5 m s$^{-1}$ with the zero line removed. Gray contours represent the 20, 30 m s$^{-1}$ isotachs of the 300 hPa mean zonal wind.
Fig. 10. Modeled mean winter cold tongue index (CTI) for each model regressed onto the modeled mean winter 300 hPa zonal wind field. Solid (dashed) black lines indicate positive (negative) perturbation isotachs, contoured every 1 m s$^{-1}$ with the zero line removed. Gray contours show the 20, 30 m s$^{-1}$ isotachs of the model mean 300 hPa zonal wind.
Fig. 11. (a) Total inter-model variance of the mean winter 300 hPa zonal wind and (b) Inter-model variance of the mean winter 300 hPa zonal wind explained by the SST expansion coefficient of the first mode of MCA covariability. Variance is expressed in units of m$^2$ s$^{-2}$ and contoured every 10 m$^2$ s$^{-2}$.
Fig. 12. Scatter plot of the normalized projection of each model’s PC/EOF 1-2 onto the corresponding reanalysis PC/EOF 1-2 for (a) the Pacific, (b) the Atlantic. EOF 1 is indicated by asterisks (*) and EOF 2 is indicated by crosses (+). Dashed circles indicate lines of constant correlation at $r = 0.25, 0.5, \text{ and } 0.75$. 
Fig. 13. Longitude of the maximum wind perturbation associated with EOF 1/2 regressed onto the longitude of the jet exit region in each model for (a) Pacific EOF 1 (Extend/Retract), (b) Pacific EOF 2 (Shift), (c) Atlantic EOF 1 (Shift), (d) Atlantic EOF 2 (Extend/Retract). Open circles represent reanalysis data and the dashed line shows $y=x$. Longitudes are displayed in degrees east (e.g. $240^\circ$E $= 120^\circ$W).