J7.4 STRATOSPHERIC CHANNELS OF ROSSBY WAVE-TRIGGERED INERTIAL INSTABILITY

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1. INTRODUCTION

Inertial instability is a hydrodynamic instability of strongly anticyclonic flow (Knox, 2003). It is characterized by anomalous potential vorticity (PV), i.e. negative PV in the Northern Hemisphere. It has been best documented in the middle atmosphere, where theory (Dunkerton, 1981; Boyd and Christidis, 1982; Griffiths, 2003a), observations (Hitchman et al., 1987; Hayashi et al., 1998) and modeling (e.g., O'Sullivan and Hitchman, 1992; Sassi et al., 1993) have established the reality of inertial instability near and above the equatorial stratopause.

This instability is physically manifested in the equatorial middle atmosphere as vertically thin "pancake structures" in temperature and divergence that can extend for tens of degrees of latitude and longitude. Research using United Kingdom Met Office (MetO) analyses and Upper Atmosphere Research Satellite (UARS) instrument data (Orsolini et al., 1997; Limpasuvan et al., 2000) has related inertial instability to the two-day wave at the equatorial stratopause and in the mesosphere. Inertial instability may also affect the development of stratospheric sudden warmings (Rosier and Lawrence, 1999), implying a role for inertial instability in middle atmosphere dynamics beyond and below the equatorial stratopause.

However, multiyear studies of inertial instability have been, to our knowledge, limited to the upper troposphere (Schumacher and Schultz, 2001; Sato and Dunkerton, 2002), the stratopause (Hayashi et al., 1998), or to selected rocketsonde sites (Hayashi et al., 2002). No long-term stratospheric climatologies using global winds have been performed, possibly owing to doubts concerning the accuracy of tropical winds in these data sets. Without such a global climatology, the significance of individual case studies can be difficult to assess: are these isolated and rare events, or are they snapshots of a more widespread and consequential phenomenon?

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Observational (Hitchman et al., 1987; Smith and Riese, 1999), theoretical (e.g., Dunkerton, 1993) and numerical (e.g., O'Sullivan and Hitchman, 1992; Clark and Haynes, 1996) studies have linked inertial instability events with Rossby wave breaking in the middle atmosphere. The Rossby wave breaking appears to organize the instability both latitudinally and longitudinally, while the preferred vertical scale of the instability is a matter of ongoing research (Griffiths, 2003b).

Here we discuss a main result of this phase of our research (Knox and Harvey, 2005): the existence of channels of inertial instability in the stratosphere that are collocated in space and time with regions of Rossby wave breaking.

2. DATA AND METHODOLOGY

MetO assimilated daily (1200 UTC) analyses of temperature, geopotential height and winds from 1 November 1991 through 31 March 2004 are used to examine the timing and pattern of occurrence of inertial instability. The analyses are generated in a manner similar to operational weather forecasting, using the technique of data assimilation with data from operational meteorological sources, such as satellites and radiosondes (Swinbank and O'Neill, 1994a). This assimilation approach holds promise for doing a better job of accurately representing stratospheric winds in the tropics than other methods. As evidence for this, both the quasi-biennial and stratopause semi-annual oscillations are evident in this data set (Swinbank and O'Neill, 1994b).

The data are available on pressure levels from 1000 hPa to 0.316 hPa with a vertical resolution of 1.6 km, on a 2.5° latitude by 3.75° longitude global grid. Potential vorticity is calculated on isentropic surfaces from the 330 K surface (approximately 13 km, just below the level of the tropopause; Knox, 1998) up to 2000 K (about 50 km, the stratopause).

Inertial instability has been diagnosed in the following manner. The condition for inertial instability criterion in a geostrophic flow (Holton, 1992) is:

$$f\left(f+\zeta_g\right)<0, \qquad (1)$$

in which f is the Coriolis parameter on an f-plane and ζ_g is the geostrophic relative vorticity. This condition is identical to the presence of anomalous PV in a statically stable geostrophic background flow. Gridpoints satisfying equation (1) are identified for each day. Monthly probability maps are created depicting the percentage of days in each month that the criterion is satisfied. These calculations are performed on potential temperature surfaces from 330 K to 2000 K. We do not impose a temporal continuity requirement since the e-folding time for the instability, $\left[-f\left(f+\zeta_g\right)\right]^{-1/2}$, can be as rapid as one day. While regions in which equation (1) is satisfied do not always guarantee the physical occurrence of inertial instability because of assumptions in the theory (Knox, 1996) and errors in the data, those regions will be referred to as being inertially unstable for the purposes of this paper.

To diagnose Rossby wave breaking, we follow Postel and Hitchman (1999) and others by calculating meridional gradients of PV and identifying regions where d(PV)/dy < 0. Fourth-order differencing and spatial smoothing are used to emphasize larger-scale wave breaking.

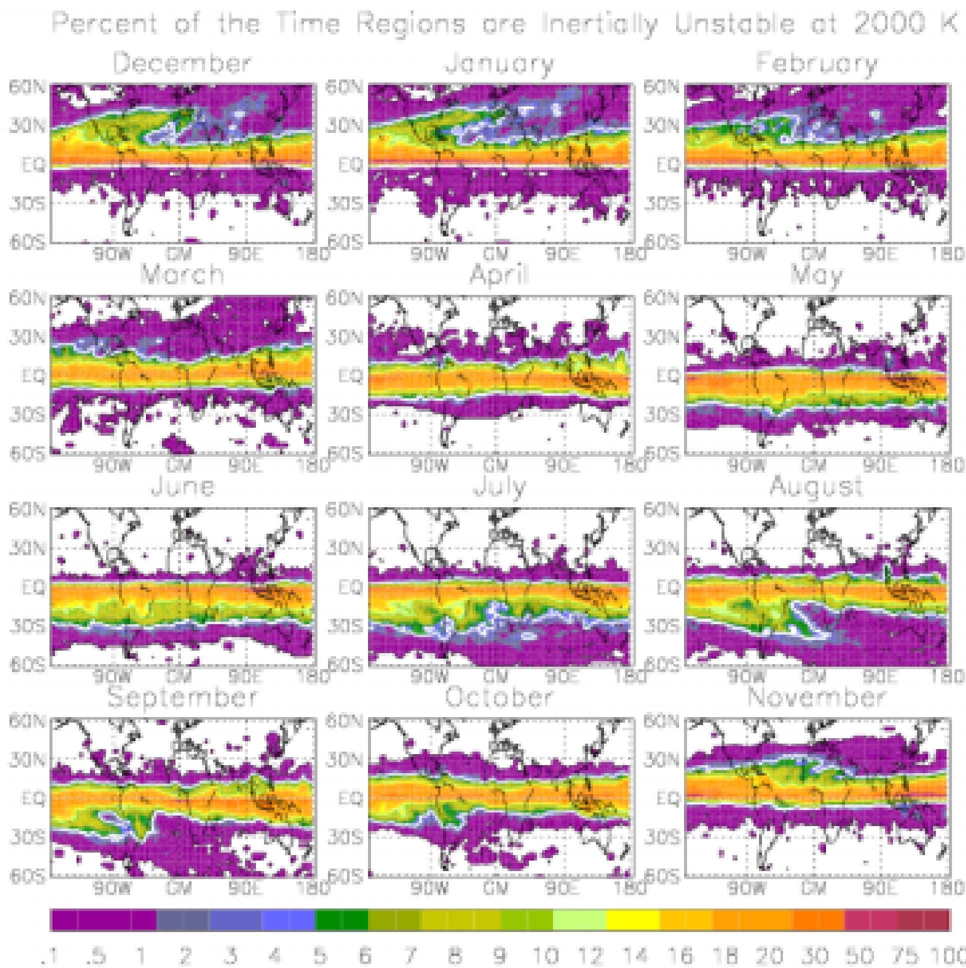

Figure 1. Monthly inertial instability frequency on the 2000 K isentropic surface for 60˚N - 60˚S during the period 1 November 1991 - 31 March 2004. The white line indicates the 5% frequency level.

3. RESULTS FOR UPPER STRATOSPHERE

Figure 1 depicts the month-by-month frequency of daily inertial instability occurrence in the MetO analyses for 1991-2004 on the 2000 K isentropic surface. The highest frequencies are in zonal bands near the equator, the center offset slightly (up to about 10˚ of latitude) into the winter hemisphere. The highest frequencies exceed 50% of days in January, which equates cumulatively to about 180 days of occurrence during all Januaries in the climatology.

This latitudinally banded pattern is to be expected, since the equatorial region is where inertial stability is lowest due to the smallness of the Coriolis parameter. However, the zonally *asymmetric* nature of inertial instability in wintertime is also clearly manifested in this figure, as discussed below.

In the Northern Hemisphere, a poleward-and-eastward "tongue" or channel of elevated (5-10% of all days) occurrences exists during the months of November through March. In Figure 1, it stretches from over the central Pacific southwest of Hawaii all the way to central Asia near 40ºN. In January, occurrences of a few percent (i.e., about twelve days over the entire climatology) extend poleward and eastward to roughly 60˚N, 135˚E, i.e. Siberia.

This tongue corresponds well in location, altitude and timing with multiyear analyses of inertial instability in the stratosphere. Hayashi et al. (1998) discuss seven equatorial "pancake structure" events in UARS Cryogenic Limb Array Etalon Spectrometer (CLAES) temperature data during the Northern Hemisphere winters of 1991-1992 and 1992-1993, five of which occur between roughly 180˚E and 270˚E—the southern terminus of the poleward-and-eastward tongue in our climatology. Hayashi et al. (2002) identify a wintertime (November-March) maximum at Kwajalein (8.7˚N, 167.7˚E) in rocketsonde observations made from 35-65 km during 1969-1993. Kwajalein is west-southwest of Hawaii, again at the southern end of the tongue of elevated inertial instability frequency seen in Figure 1.

Individual case studies of near-stratopause inertial instability also correlate well in space and time with the results in Figure 1:

- Hayashi et al. (2002) include case studies from December 1978 and February 1979, drawing from the work of Hitchman et al. (1987), who also discuss a case in January 1979. Latitudinally, all three cases are confined between approximately 0-20˚N; longitudinally the first two are centered near Kwajalein, and the January 1979 case appears centered farther east, near 120˚W. These locations correlate well with the poleward-eastward tongue of highest inertial instability frequency in Figure 1, and the timing correlates well with the November-March maximum in this tongue feature.

- The case of inertial instability in Cryogenic Infrared Spectrometers and Telescopes for the Atmosphere (CRISTA) data studied by Smith and Riese (1999) occurred in the upper stratosphere in early November 1994 between roughly the latitudes of 0-30˚N (with pronounced amplitudes in the 25-35˚N range) and a longitude band of 120˚W-90˚E and extending eastward during the period. Again, this location is consistent with the elevated tongue of inertial instability frequency seen during November in Figure 1. The more poleward location of this episode versus the cases examined by Hayashi et al. (2002) and Hitchman et al. (1987) is explained simply by noting that the CRISTA case happened to occur farther east than the others along the same poleward-and-eastward channel. Smith and Riese (1999) also note in their Figure 4 the presence of inertially unstable "pancake structures" at the end of the episode near 30-40˚N, 30-90˚E—which once again is located farther along the poleward and eastward channel in our Figure 1.

- Orsolini et al. (1997) discuss stratopause inertial instabilities in MetO analyses in mid-January 1992 that occur near 10-30˚N, 100-225˚E, with a much broader region of low PV at approximately the same latitudes but in the Western Hemisphere at 250-315˚E longitude. The narrowly longitudinally confined instabilities do not occur along the poleward-eastward channel, but the broad low-PV region matches the channel location quite well. The timing of inertial eddy episodes at the stratopause in 1991-1995 in Figure 3 of Orsolini et al. (1997) also appears to agree with our climatology, occurring episodically from mid-December through mid-February.

This poleward-and-eastward tongue in our inertial instability climatology is mirrored in the results for austral winter (June-September), when a narrow maximum of inertial instability occurrence (up to about 10%) extends from the central equatorial Pacific to the mid-latitudes between Africa and Antarctica. This mirror-image extending to mid-latitudes in both winter hemispheres supports the contention that the inertial instability frequency patterns are physical, and are not spurious features due to data assimilation errors in tropical winds.

However, the key to establishing the physical reality of the inertial instability features is to relate them in space and time to Rossby wave breaking, which presumably triggers the inertial instability. Therefore, in Figure 2 we juxtapose the inertial instability frequencies at 2000 K with Rossby wave breaking frequency. Because of the seasonal dependence of inertial instability and Rossby wave breaking, results are shown for December-January in the Northern Hemisphere and August-September in the Southern Hemisphere. For clarity, only Rossby wave breaking frequency above 25% is depicted, via stippling.

The regions of elevated inertial instability at 2000 K are almost exactly coincident in space with the regions of Rossby wave breaking, the wave breaking criterion being satisfied just a few degrees of latitude (one or two gridpoints) equatorward of the maxima in inertial instability frequency. This is true in the Northern Hemisphere from near Hawaii poleward and eastward to Europe, and in the Southern Hemisphere from near

Samoa at 15˚S, 180˚ and then poleward and eastward to south of the Cape of Good Hope.

The wintertime inertial instability frequencies in Figure 2 can be compared to Hayashi et al.'s (2002) results for Kwajalein, where frequencies in rocketsonde data from 35-65 km ranged from 8-12% for December-February. Our climatology indicates wintertime inertial instability frequencies near 20% at Kwajalein, but our results in Figure 2 are for the stratopause and decline to less than 10% at Kwajalein in the middle stratosphere (not shown). Given these considerations, the agreement in inertial instability frequency between our climatology and Hayashi et al. (2002) is excellent.

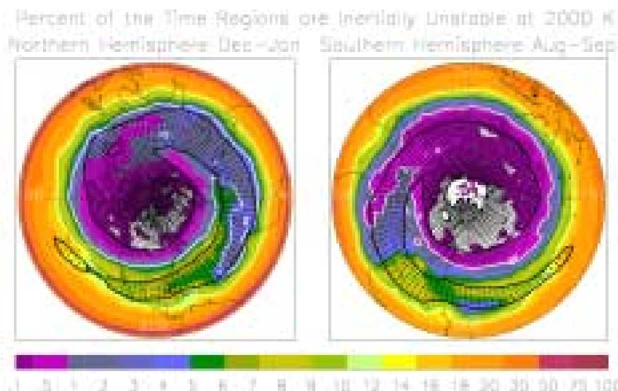

Figure 2. Polar stereographic graph of mid-winter inertial instability frequency (shading) and highest Rossby wave breaking frequencies (frequencies above 25% stippled) in the Northern and Southern Hemispheres at 2000 K during the period 1 November 1991 - 31 March 2004. The white line indicates the 5% frequency level in the Northern Hemisphere and 2% in the Southern Hemisphere.

On the strength of this and other evidence discussed elsewhere (see Knox and Harvey, 2005), we conclude that our climatology is indeed capturing realistic inertial instability events.

4. MIDDLE STRATOSPHERE RESULTS

The same patterns in inertial instability frequency and Rossby wave breaking found at 2000 K exist, albeit with diminished frequencies, downward into the middle stratosphere. The spatial continuity between these patterns at different altitudes provides indirect evidence that these results are physical, not the artifacts of data assimilation, and that the instability is triggered by deep atmospheric structures such as Rossby waves.

It is of interest to investigate the interannual variability of these features. As a first step, Figure 3 depicts the year-to-year frequencies of inertial instability and Rossby wave breaking at 1000 K over the Caribbean Sea, in the poleward-and-eastward maximum. This is done by calculating *maximum* frequencies for each winter month anywhere within the box in Figure 3, and then averaging across the entire winter season.

The results in Figure 3 show that inertial instability is not rare, particularly over the Caribbean where the instability criterion may be satisfied up to a maximum of 50% of the time in winter. In addition, the interannual variabilities of inertial instability and Rossby wave breaking are broadly correlated at both locations, as expected. Our climatology permits us to quantify the correlations.

For the Caribbean "hot spot" identified in Figure 3, the correlation coefficient r is 0.58, and r = 0.50 for the zonal mean at the latitude of the "hot spot," 16.25˚N. Using the Fisher transformation (Wilks, 1995), these correlations are significant at the 5% level for the "hot spot," and at the 10% level for the zonal mean. (In the Southern Hemisphere [not shown], the correlations are lower, perhaps due to the weaker Rossby wave signal there: r = 0.32 for the austral "hot spot" and r = 0.22 in the zonal mean analysis at 18.75˚S.)

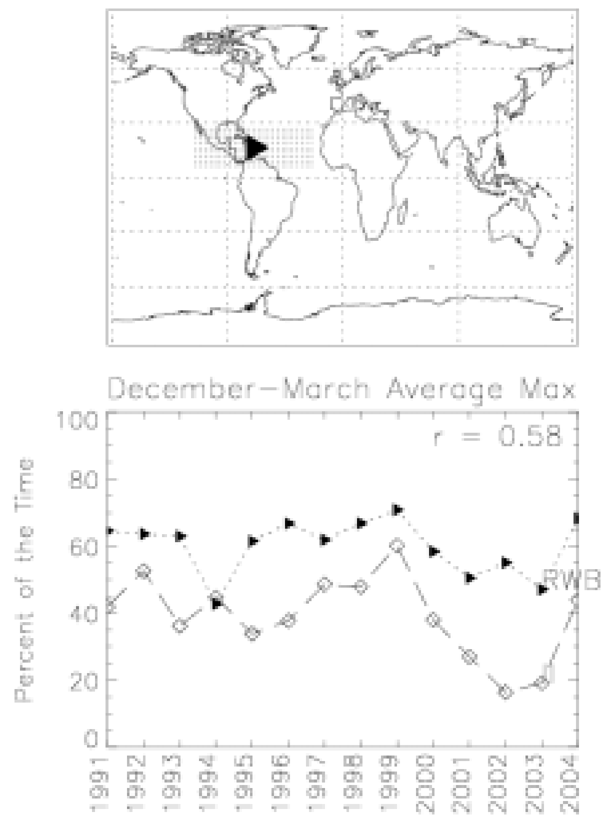

Figure 3. Time series of winter-season averages of monthly maximum wintertime frequency of Rossby wave breaking (RWB, solid triangles in lower graph) and inertial instability (II, open diamonds in lower graph) at 1000 K for the Caribbean and adjacent areas (indicated by stippled box at top, centroid at 16.25˚N, 290.625˚E).

The seesaw in inertial instability frequency over the Caribbean in the early-to-mid-1990s is possibly related to the quasi-biennial oscillation (QBO), as hypothesized by Knox (1996, pp. 228-229). Easterly QBO events in the middle stratosphere, such as occurred during the winters of 1991-92, 1992-93, and 1994-95, provide greater meridional shear in tropical latitudes that could facilitate the development of inertially unstable regions. As seen in Figure 3, these winters did in fact possess more inertially unstable days than did the westerly-QBO-phase winters of 1993-94 and 1995-96. The impact of wind oscillations on the long-term variability of Rossby wave-triggered inertial instability is the subject of future work.

However, the most prominent feature of interannual variability in the later years of the climatology in Figure 3 is the marked downturn in both inertial instability and Rossby wave breaking frequencies over the Caribbean in 2001-2003. The reasons for this downturn and the absence of a possibly QBO-induced signal during this period are not understood and also warrant additional study.

5. CONCLUSIONS AND DISCUSSION

The primary intent of this study is to use a long-term global climatology of inertial instability and Rossby wave breaking to extend our understanding of inertial instability and make connections between previous research efforts.

We have found that inertial instability is neither rare nor confined to near the equatorial stratopause. Instead, in the upper stratosphere it occurs roughly 5-10% of the time in poleward-and-eastward “tongues” or channels extending from the equatorial regions to the mid-latitudes in both winter hemispheres. These channels are deep and extend, with decreasing frequency, down into the middle stratosphere. The Northern (Southern) winter channel extends from the equatorial Pacific to Asia (South Africa).

According to our climatology, the preferred longitude region for inertial instability depends on the latitude of the breaking, with more poleward events occurring farther eastward along the poleward-and-eastward channel. All of these channels are just poleward of maxima in Rossby wave breaking frequency, further reinforcing the intimate connection between inertial instability and Rossby wave breaking in the middle atmosphere.

Nearly all previous studies of inertial instability have identified cases of the instability that lie in or near these channels. Future research on inertial instability in the middle atmosphere can and should be targeted fairly narrowly in space and time, because the events appear to be focused rather tightly on specific regions and seasons.

Similarly, tracer transport due to Rossby wave-triggered inertial instability should also be confined to tightly prescribed regions along the poleward-and-eastward corridor of enhanced instability frequency. Smith and Riese (1999) examined the possible transport of methane and ozone by inertially unstable circulations, but did not explore the possibility that the transport could be narrowly delineated in time and space.

It should be emphasized that these regions are not latitude bands, but instead extend poleward and eastward from near the equator. Therefore, latitudinal averages would obscure the impact of inertial instability on tracer transport, whereas an average downstream along one of the poleward-and-eastward channels might more accurately capture the impact of the instability on tracers.

The poleward extent of the “tongues” raises the possibility that inertial instability can interact with the dynamics of the mid-latitude stratosphere, and beyond. We intend to explore the two-way interaction between inertial instability and extratropical dynamics in future work.

The chronic juxtaposition of inertial instability nearby, but poleward of, Rossby wave breaking maxima appears to be related to the spatial pattern of Rossby wave breaking (Peters and Waugh, 1996). These breaking events are at low latitudes and appear to maximize about a month prior to the overall maximum in Rossby wave amplitude in both hemispheres. In a low-latitude Rossby wave breaking event, the anomalous PV extruded from the opposite hemisphere would, as a result of its folding-over, satisfy the Rossby wave breaking criterion used in this study at a point equatorward of the anomalous PV. In a “P2” breaking event in the Peters and Waugh classification scheme, the region of anomalous PV poleward of the breaking is broad and conducive to vigorous inertial instability. The frequency and nature of these events would be modulated by the background zonal-mean zonal shear induced by the SAO and/or QBO, which may also contribute to the early-season maximum in events relative to the overall Rossby wave amplitude maximum.

Finally, the robustness of the spatial patterns in both hemispheres for a long-term climatology, combined with the corroboration of extant case studies and climatologies, strongly suggests that these results are not the product of bad data or poor data assimilation. Inertial instability studies have long suffered from doubts regarding data quality, since the instability tends to occur in regions where the data is sparse or poor, and where dynamical assumptions allowing wind data to be inferred indirectly are least reliable. This study, which spans over a decade and links indirect indices of inertial instability occurrence with direct observations of “pancake structures” and tracer transport in individual case studies, provides persuasive evidence of the reality of this instability.

6. ACKNOWLEDGMENTS

This work is partially supported by NASA UARS Science Investigation NRA (Contract NAS5-01158) via NorthWest Research Associates subcontract NWRA-01-S-060. Thanks to Ruth Lieberman of CoRA/NWRA for making this support possible. We would also like to thank colleagues at the Met Office for producing the stratospheric assimilated data set. We are grateful to the British Atmospheric Data Centre for providing us with access to the MetO data. This paper is a condensed, revised version of Knox and Harvey (2005).

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