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The late spring Caribbean rain-belt: climatology and dynamics

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ABSTRACT: This study examines the dynamics of late spring rainfall (the Early Rainy Season, ERS) in the Caribbean region, in hopes of identifying mechanistic-based predictors for low-frequency climate modulations of the system. The subtropical Caribbean rain-belt develops in May as seasonal warming proceeds. By July, the rain-belt retreats north apparently following the westerlies and their vigorous synoptic disturbances. Daily climatology data suggest a physical definition of the Caribbean ERS as mid-May to mid-late June. Based on an examination of daily loops for several seasons, we hypothesize that rainfall occurs quasi-randomly throughout tongues of air with sufficiently high (above 45–50 mm) precipitable water (PW). These moist airmasses are brought north from the deep tropics by low-level southerlies, and typically bent over into SW-NE bands by latitudinal shear of the westerlies. The low-level flow that transports PW tongues is partly induced by upper-level synoptic disturbances in the westerlies, but also involves the gentle persistent flow around a geographically anchored Panama Low. While forced ascent is sometimes active ahead of these upper-level troughs, convective and mesoscale processes can produce rain wherever PW is sufficient. In summary, we hypothesize that rainfall hinges largely on *the Lagrangian statistics of moist air tongues*. Comparison is drawn between the Caribbean rain-belt and its East Asian counterpart (Meiyu-Baiu), and other mechanisms and diagnostics from that literature are discussed. Statistical prediction exercises, based on mechanistically chosen predictors, could both test hypotheses and aid local agricultural interests in the region.

KEY WORDS   Caribbean precipitation; Caribbean rain-belt; mid-summer drought; Caribbean early rainfall season; Caribbean; precipitation

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1. Introduction

Rainfall during the early rainy season (ERS) is critical for Caribbean agriculture during the early growing season (Gamble *et al.*, 2010). Reduced rainfall after the ERS results in a decline in vegetation health during the mid-summer (Allen *et al.*, 2010). Definitions of the ERS vary, sometimes April to July, or May to July (Spence *et al.*, 2004), or May to June (Chen and Taylor, 2002). Our goal here is to clarify thermodynamic and dynamical mechanisms driving the ERS, in order to choose physically based statistical predictors for year-to-year variability.

Thermodynamically, ERS onset may depend on crossing of a sea surface temperature (SST) threshold for deep convection, about 26–27 °C in the tropics depending upon region and season (Waliser and Gautier, 1993). The Atlantic Warm Pool (AWP), a boreal summer warm water feature, satisfies the SST criterion for deep convection in the Caribbean. Evaporation from warm water can help to elevate column integrated water vapour or total precipitable water (TPW), which may exert the more direct or proximate control on convection. Still, whether direct or indirect, the dependence of rainfall on SST helps motivate and explain interannual variability. Previous studies suggest that Caribbean rainfall variations may be driven by fluctuations in local SST (Curtis, 2013), and AWP onset date exhibits similar variability with ERS onset dates (Misra *et al.*, 2014).

Dynamical mechanisms also appear to play a role in governing rainfall, with implications for its interannual variability. Regional climate involves open systems, subject to external influences as well as internal sources of variability. Understanding the basic system of interest (in this case, a rain-producing weather pattern) is not only of intrinsic interest, but also can help guide the practical search for associations with possibly predictable remote or large-scale influences. One path towards such an understanding is through deduction from the basic facts of mean seasonality. Another path involves interpretation of samples of the actual instantaneous weather that makes rain.

In this study, we show that the Caribbean late-spring ERS is best defined physically as mid-May to mid-June, rather than in calendar monthly mean averages. We hypothesize that ERS onset may be governed by seasonal warming and moisture availability, while its demise (the onset of July's drier conditions) may be more dynamical, tied to the retreat of the midlatitude westerlies and their troughs, in conjunction with the enhancement of easterly flow

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at all levels, and attendant enhanced low-level easterly trades on the flank of the dry Atlantic Subtropical High. These hypotheses can guide research into the causes and predictability of ERS interannual variability, which could both aid regional interests and also serve as an arena for testing the hypotheses themselves.

Towards these goals, this paper reviews the mean spring–summer climatology of the region (Section 3), examines synoptic processes during the ERS, summarized in a schematic (Section 4), and makes comparisons to the better-studied analogue of East Asia (Section 5). Conclusions and prospects for increasing understanding further are considered in Section 6.

2. Data sources

For this study, gridded precipitation estimates covering land and water are favoured over station observations, since regional patterns are our subject, not quantitative values on rugged heterogeneous terrain. Principally we use the 0.25° Tropical Rainfall Measuring Mission 3B42 (Huffman *et al.*, 2007), a three-hourly original data product examined here in daily means from 1998 to 2012. Some figures also use GPCP (Adler *et al.*, 2003) or interpolated outgoing longwave radiation (OLR, Liebmann and Smith, 1996) to characterize cloud and rain patterns, as mentioned in captions and text. The Modern Era Retrospective Analysis for Research and Applications dataset (MERRA, Rienecker *et al.*, 2011) was used to analyse moisture (TPW) and other fields, along with the older NCEP-NCAR Reanalysis I (Kalnay *et al.*, 1996) for some meteorological fields. None of our mostly qualitative conclusions hinge on the modest quantitative differences among these gridded data products and reanalyses.

3. Caribbean region summer rainfall

In this section, mean climatology is shown and described (Section 3.1), followed by a brief review of studies of low-frequency modulations and how they relate to the basic phenomenology (Section 3.2).

3.1. Climatological mean and seasonal patterns

The summer mean rainfall pattern in the region is shown in Figure 1. Local features related to continents, islands, peninsulas, and perhaps the Gulf Stream are superposed on a very broad scale underlying SW-NE tilted zone of rainfall. However, this seasonal accumulation pattern blurs together distinct subseasonal regimes, as we shall see.

The annual rainfall cycle over the Caribbean basin reveals a distinct bimodal pattern, with peaks during the late-spring and late-summer months, as seen in Figure 2(a)'s climatological (January–December) GPCP rainfall time series in each grid cell of the region. In many places, a relative minimum during the mid-summer, known as the mid-summer drought (MSD) (Magaña *et al.*, 1999), separates the *Primera* or the ERS from the late-rainfall season (LRS) precipitation peak. During this midsummer period, some of the farthest north boxes in Figure 2(a) have their peak rain. The complementary timing of these events (emphasized by the larger area averaged red *vs* black series in Figure 2(b)) suggests that a moving rain-belt, retreating into middle latitudes along with the summer jet stream, may be a more relevant description of the onsets and demises of sub-season features in local time series. Karnauskas *et al.* (2013) offer related observations and interpretations. Following the MSD, a second and longer-lasting (see Figure 2(b)) rainfall peak in the Caribbean during the LRS coincides with the warmest SST and a higher frequency of passing easterly waves and tropical storms.

As an aside, the influence of the Gulf Stream on the mid-latitude mid-summer rainfall has been considered in modelling studies (Minobe *et al.*, 2008; Kirtman *et al.*, 2012). However, results suggest that a climatological rain-belt in those latitudes exists even in the presence of a smoothed SST field that eliminates the pattern of the warm Gulf Stream (Yoshida *et al.*, 2010).

The sub-seasonal pattern evolution is clearer in Figure 3, showing climatological mean rainfall, sea level pressure, low-level winds, low-level and upper-level geopotential heights, and SST for April (top), the ERS (middle), and July (bottom). In April the Caribbean is under westerlies (Z250 contours) but apparently is still too cool for abundant convection and rainfall (SST). In July, Caribbean conditions are warm and rain occurs over the islands, but the upper-level westerlies (and thus their disturbances) are confined north of Florida. Only in the ERS does the rainfall span ocean and land across the Greater Antilles. A distinct dry zone through the Gulf of Mexico is seen during the ERS, which suggests that simple west-to-east motion of rain-producing mid-latitude fronts along the mean westerlies cannot explain the pattern in the Caribbean. This distinct belt will hereafter be referred to as the *Caribbean rain-belt*. On average, the Caribbean rain-belt extends diagonally from the Southwestern Caribbean across the Greater Antilles and into The Bahamas along approximately 13°–24°N.

A southerly flow component at low levels (implying a flux of tropical moisture into the region) is also prominent in the ERS, while the flow is more zonal in April and July towards the western Caribbean. Specifically, the 850 hPa meridional wind time series (Figure 4) gives another clear indicator for defining the ERS from mid-May to mid-June, and may be a response to the latent heating arising from condensation throughout the rain-belt (as suggested by the model calculations in the Appendix of Sampe and Xie, 2010), not an external cause. Besides following the flank of the North Atlantic Subtropical High (NASH), geostrophic flow at low levels is also affected by a smaller-scale heat low (a mild trough in the pressures and heights) near the Caribbean coasts of Panama and Colombia, known locally as the Panama Low. A region of deep convection and moisture convergence regularly occurs around the Panama Low, which contributes to the formation of meso-scale convective clusters (Poveda and Mesa, 2000).
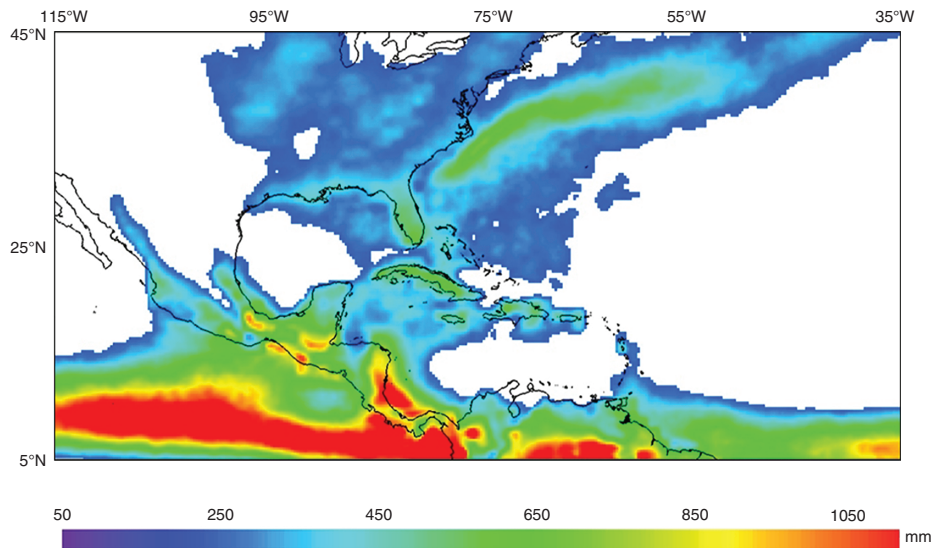
Figure 1. Average accumulated rainfall (mm) between 15 May and 31 July (1998–2012, TRMM 3B42).

During July, surface pressure in the western Caribbean has a 'bump' described by Curtis and Gamble (2008) and Mapes *et al.* (2005) during the MSD, as the NASH extends further into the western Caribbean. This may contribute to the drying during the MSD period (Kelly and Mapes, 2011).

3.2. Low-frequency (interannual) modulations

Recall that a goal in describing the climatology of the Caribbean rain-belt system was to seek influences that may also explain year-to-year differences. Interannual variability is observed in annual rainfall totals, length of the rainy season and length and timing of the MSD (Magaña *et al.*, 1999; Jury *et al.*, 2007; Curtis and Gamble, 2008; Gamble *et al.*, 2008; Angeles *et al.*, 2010). Various mechanisms have been proposed to explain the MSD, including regional SST anomalies (Magaña *et al.*, 1999), a westward extension of the NASH (Giannini *et al.*, 2000; Kelly and Mapes, 2011), variations in the summertime Caribbean Low-Level Jet (Muñoz *et al.*, 2008), contributions from Saharan dust (Angeles *et al.*, 2010), and the twice-a-year crossing of the solar declination (Karnauskas *et al.*, 2013). However, the dynamics responsible for the onset and decay of rainfall in the Caribbean during the ERS along with its associated variability remain poorly understood. It is possible that a modified explanation of the MSD, cast as the *cessation (or northward displacement) of* the Caribbean rain-belt, may provide an additional way of understanding the mid-summer drying.

The El Nino-Southern Oscillation (ENSO) phenomenon is known to affect rainfall in the Caribbean region (Giannini *et al.*, 2001; Chen and Taylor, 2002). Early season rainfall is reduced prior to a mature El Nino and is enhanced following a mature El Nino and during La Nina years. Drier than normal summertime conditions are exacerbated when a positive winter NAO phase is observed immediately prior to a developing El Nino (Giannini *et al.*, 2001). The mechanism of ENSO and NAO influences appears to involve anomalous surface divergence and windspeed-induced SST anomalies in the Caribbean. But for any rain-producing weather system, warmer SST may also tend to increase rainfall through Clausius–Clapeyron effects alone, so a sensitivity to local or regional SST is not very discriminating about mechanisms. Dynamical mechanisms, ranging from the local tendency of warmer air, to rise to highly nonlocal effects from Asia via planetary-scale momentum fluxes (Kelly and Mapes, 2011, 2013), may contribute additional potentially predictable signals to Caribbean rainfall.

4. A case study, schematic, and new hypothesis

A positive relationship is seen between rainfall and upper tropospheric zonal wind (which steers disturbances) and rainfall in the Caribbean during the early summer months (Maloney and Esbensen, 2007; Small *et al.*, 2007). More specifically, troughs steered by upper-level the westerlies are known to cause Caribbean wet spells (Jury *et al.*, 2007), in agreement with SX10 Meiyu-Baiu conclusions. Does this imply that QG forcing for uplift, such as ahead of troughs due to positive vorticity advection aloft, is the driver? Not necessarily: Low-level manifestations of troughs include low-level cyclones with poleward flow from the tropics on their east side. Such flow transports moisture polewards, and broadly is directed up the mean equator-to-pole slope of mean isentropes (implying isentropic upglide, discussed further in Section 5). In any one case, or in blurred averages, the effects of dynamical lifting and northward moisture transport are therefore difficult to tease apart. The distinction may not matter for some purposes, but is worth drawing nonetheless since other flow components are also active.

Consider the situation on 23 May 2009, depicted in Figure 5. This image was selected, after viewing analogous imagery loops from multiple ERS seasons, for
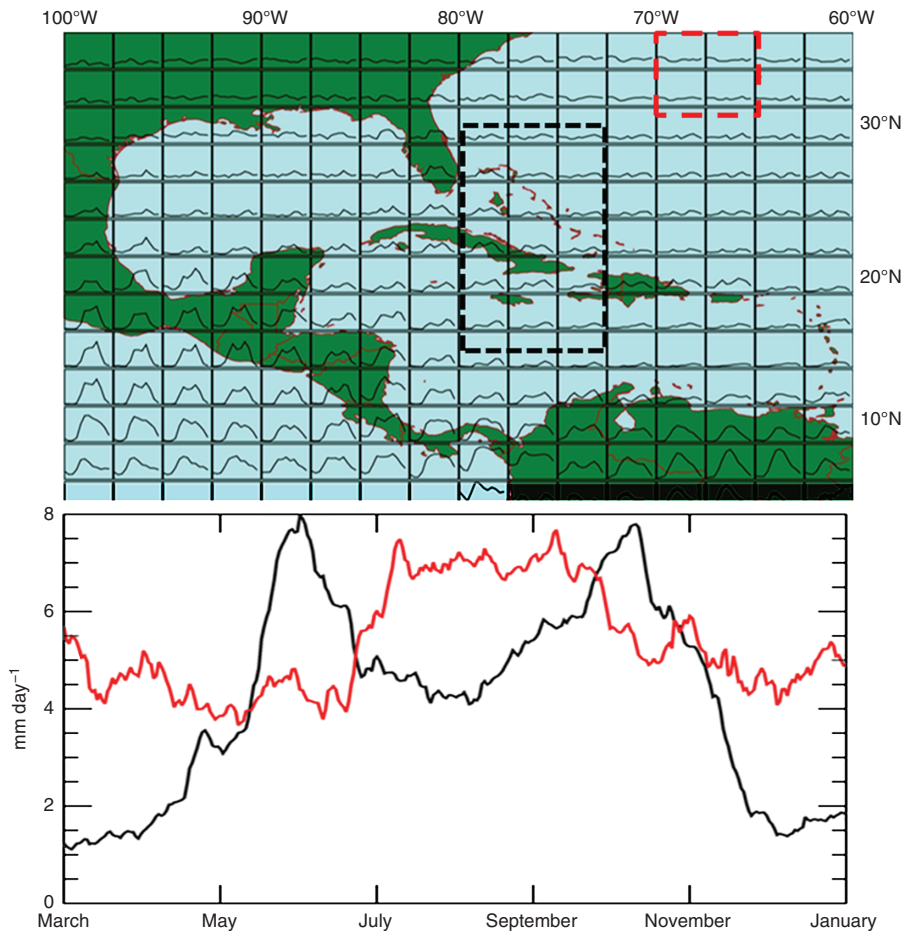
Figure 2. (a) Annual time series of climatological GPCP monthly rainfall in each 2.5° grid cell from the 1979–2012 period. Black and red dashed boxes represent the location of the Caribbean Sea and North Atlantic time series used in Figure 2(b). Notice the sharp onsets in late spring and double peak in the warm season. (b) Daily climatological time series of rainfall over the Caribbean Sea (14°–29°N, 80°–75°W, black) and the western North Atlantic (30°–35°N, 70°–65°W, red) from TRMM 3B42, 1998–2012. Based on this series, we define mid-May to mid-June as the Caribbean Early Rain Season (ERS).

being illustrative of the moisture *versus* dynamical forcing distinction. However, our subjective judgement is that it is not atypical of the varied weather situations making up the mean Caribbean rain-belt pattern. Two upper-level troughs in the westerlies (black height contours of 250 hPa height) are inducing southerly flow in the lower-level flow (red arrows are short trajectories at 850 hPa) near the Panama Low region. These instantaneously indicated flow features have also existed at prior times, and thereby have displaced moist air columns (tongues in blue shading) northwards from their source in the humid deep tropics.

Substantial daily rainfall accumulations (magenta shading) are certainly seen in the strongly forced (lifted) frontal zone or warm conveyor belt to the east of the eastern trough and its cutoff. However, in addition to that strongly forced uplift band, rainfall is also scattered throughout the entire moist zone, which has been deformed into a complex shape by the various flow disturbances. In particular, the rainfall north of Puerto Rico and Hispaniola island, or southwest of Jamaica, are in upper-level ridge or straight-flow conditions that are not indicative of strong dynamical lifting.

Figure 5 was simply one sample map view of the rather close statistical relationship between TWP and precipitation (Bretherton *et al.*, 2004; Peters and Neelin, 2006). In this fairly cool area of the tropics, a TPW exceeding 45–50 mm (as emphasized by the colour table) is a threshold-like predictor of rainfall, a necessary but not sufficient condition. The western Caribbean is notable for having high TPW yet little rainfall, a commonly observed anomaly in the region that deserves further study.

Animations of tongues of humid air (such as the lucid MIMIC-TPW imagery product, whose site http://tropic.ssec.wisc.edu/real-time/mimic-tpw/natl/main.html has archives of 72-hour loops) suggest that these are bent northeastwards by the increase of westerly flow with latitude. We hypothesize here that if rainfall occurs somewhat randomly within them, the SW-NE shape of the time-mean Caribbean rain-belt of the ERS may be explained as a *Lagrangian statistical property of tracers* (*moisture*) *in the low-level flow*. While the ultimate cause of that low-level flow is often induced circulation from upper-level potential vorticity filaments (tentacles of the polar vortex, broadly speaking), *the proximate variable that shapes ERS rainfall is hypothesized to be the shape of TPW*
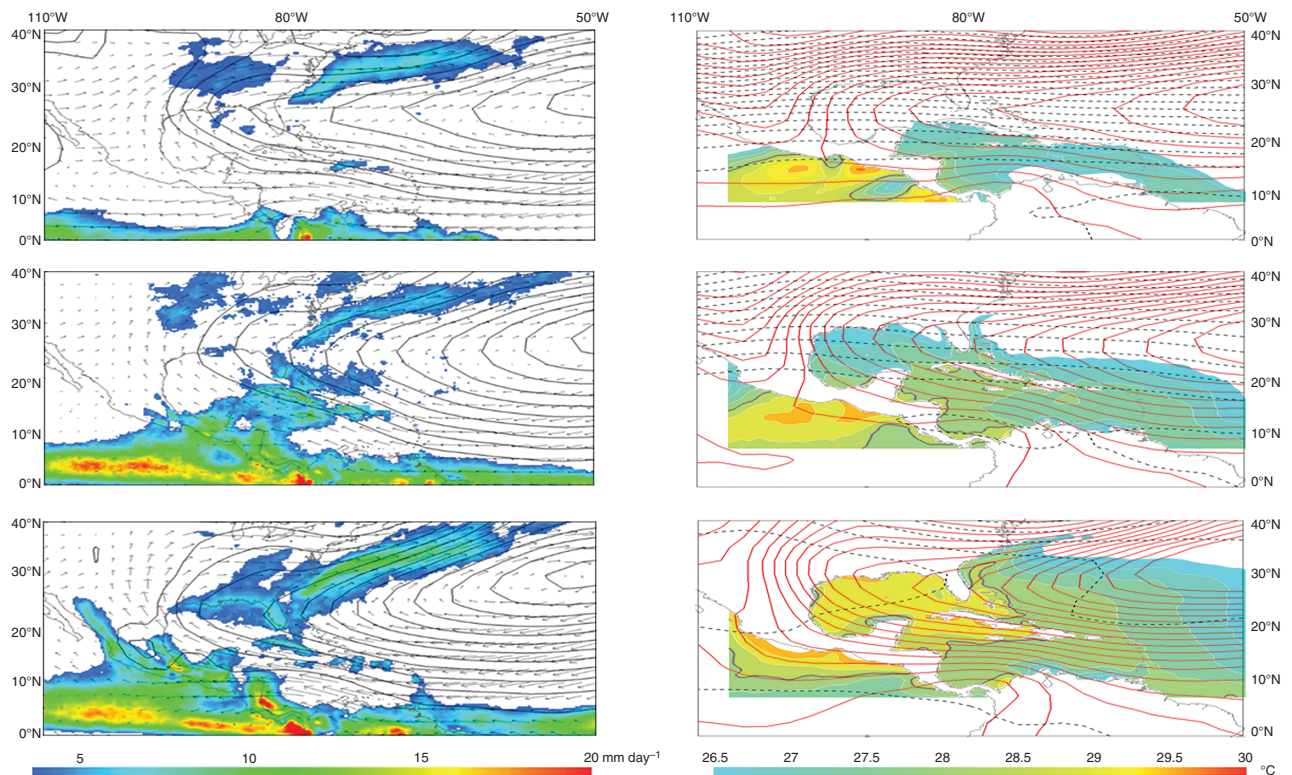
Figure 3. (a, c, e) Average sea level pressure isobars (1 hPa contours >1014 hPa), 850 hPa wind vectors (max 10 m s^{-1}), and TRMM 3B42 rain rate (shaded) for April (a), Early Rainfall Season (c), and July (e). Average baseline period calculated from 1998 to 2012. (b, d, f) Average SST (shaded > 26.5 °C) and geopotential height for 850 hPa (red contour) and 250 hPa (black dashed contour) for April (b), Early Rain Season (d), and July (f). The 26.5 °C isotherm is between the 26 and 27 °C threshold for deep convection in the tropics. Atlantic Warm Pool isotherm (28.5 °C) is indicated by a thick grey contour.

tongues. These tongues are necessarily advected by winds mainly in the lower half of the troposphere (see Mapes, 2000 for a review of tropical vertical structure issues), and can also form from the interaction of extra-tropical eddies and tropical diabatic heating (Park *et al.*, 2015).

The Lagrangian properties of the low-level flow field that shapes the TPW field are not easy to foresee from Eulerian snapshots. Steady or slowly moving features have special influence in displacing tracers. For instance, many studies in the region emphasize a quasi-steady 925 hPa Caribbean Low Level Jet (CLLJ) (<-10 m s^{-1}) during the mid-summer. There has been considerable research linking variability in the CLLJ with Caribbean rainfall anomalies (Whyte *et al.*, 2007; Muñoz *et al.*, 2008; Martin and Schumacher, 2011; Taylor *et al.*, 2013), but isotachs (which define 'jets') have no real place in theory. Instead, perhaps Lagrangian measures of the probability or frequency of occurrence of high TPW values provides a more satisfying (if harder to summarize) way to link precipitation features to flow features, favouring any combination of wide, strong, or quasi-stationary features.

Summarizing the lessons of Figures 3 and 5 and the many imagery loops we have examined, Figure 6 offers a schematic view of the climatological ERS (top) and its underlying dynamics, echoing the visual style of Figure 5 and sketched also from other data-based case studies. In this view, transients play a first-order role. Upper-level troughs (black height contours) drive unsteady low-level

flows (red arrows) that advect humid air columns, within which rainfall (pink) occurs somewhat randomly, not merely in dynamically forced regions. Low-level troughs flow associated with the quasi-stationary Panama Low also supports transport of humid air from the tropics at low levels.

5. Comparison to a well-studied analogue: Meiyu-Baiu or East Asian summer monsoon

To enlarge our pool of candidate explanations for ERS dynamics, we turn to analogues in other longitudes. Figure 7 shows ERS minus July rainfall, in a domain stretched to include Asia as well as the Atlantic. North–south rainfall dipoles observed between the South China Sea and the Sea of Japan (approximately 26°–36°N, 110°–130°E) show that in Asia too the subtropical (and tropical) rainfall bands move northwards from ERS season to July. Similar poleward-eastward tilting subtropical rain-belt features are seen on the western edge of all the subtropical anticyclones (Kodama, 1992). These climatological features have various names, both region-specific and generic, such as subtropical convergence zones (STCZ) or precipitation zones; quasi-stationary rain-belts, or subtropical frontal zones (Ninomiya, 1984; Kodama, 1992; Enomoto *et al.*, 2003; Sampe and Xie, 2010). Here we will use the term *quasi-stationary rain-belt*,
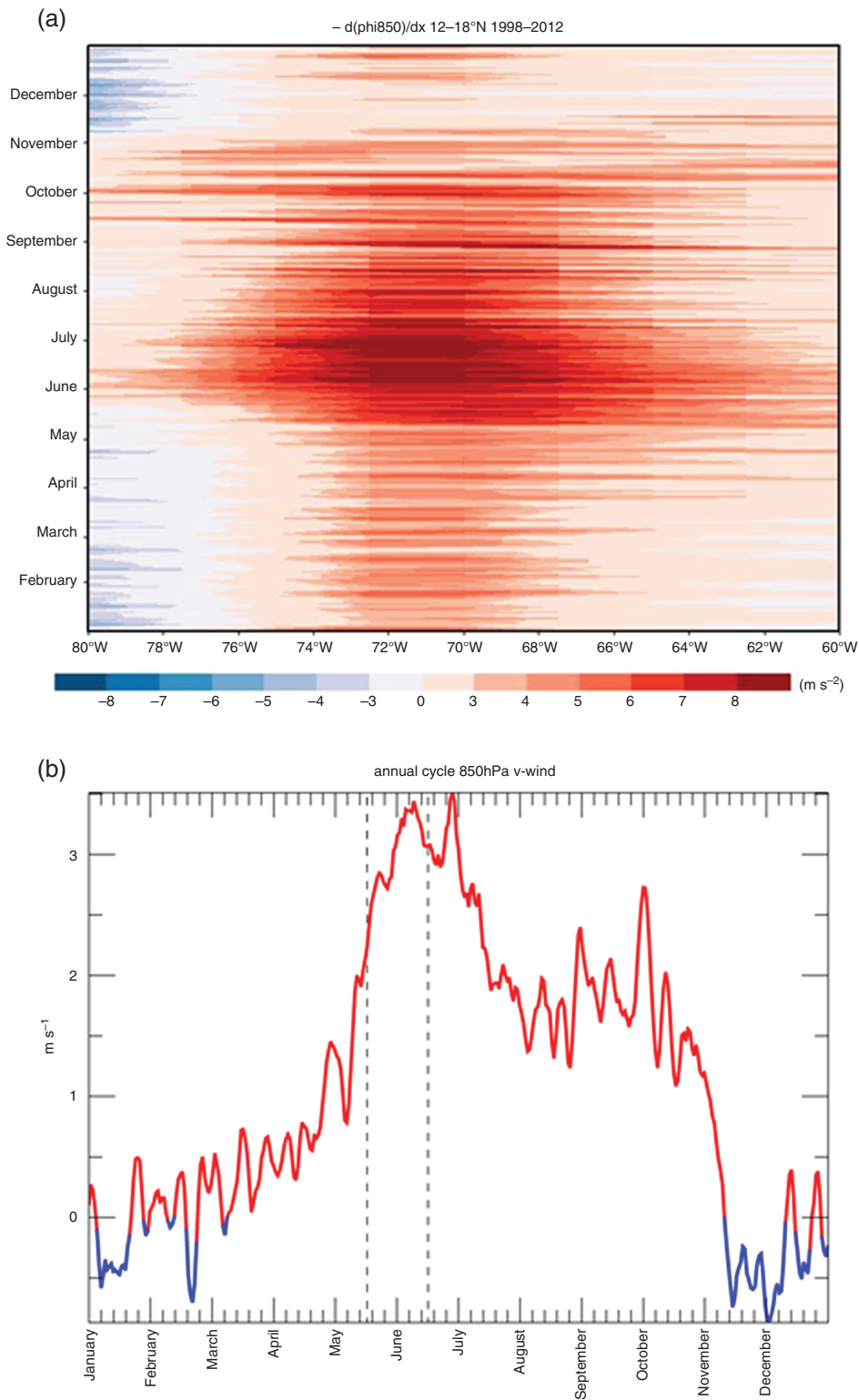
(a)

(b)

Figure 4. (a) Annual cycle of 850 hPa geopotential zonal height gradients meriodionally averaged from 12 to 18°N. Average baseline period calculated from 1998 to 2012. (b) Time series of area averaged meridional wind (red > 0 m s⁻¹, blue < 0 m s⁻¹) over the Caribbean region. Dashed vertical lines indicate the Early Rain Season period. Average baseline period calculated from 1998 to 2012.

where quasi-stationary applies on the monthly or seasonal timescale (not necessarily from day to day).

The East Asian early summer rain-belt is called *Meiyu-Baiu*, hyphenating Chinese and Japanese names (Ninomiya and Shibagaki, 2003). This rain-belt first appears near the South China Sea around mid-May (the *Meiyu* season) and migrates northwards towards Japan during the *Baiu* season where it abruptly ends in mid-July. Its climatological mean footprint is sometimes called the East Asian Summer Monsoon (EASM), and can be placed
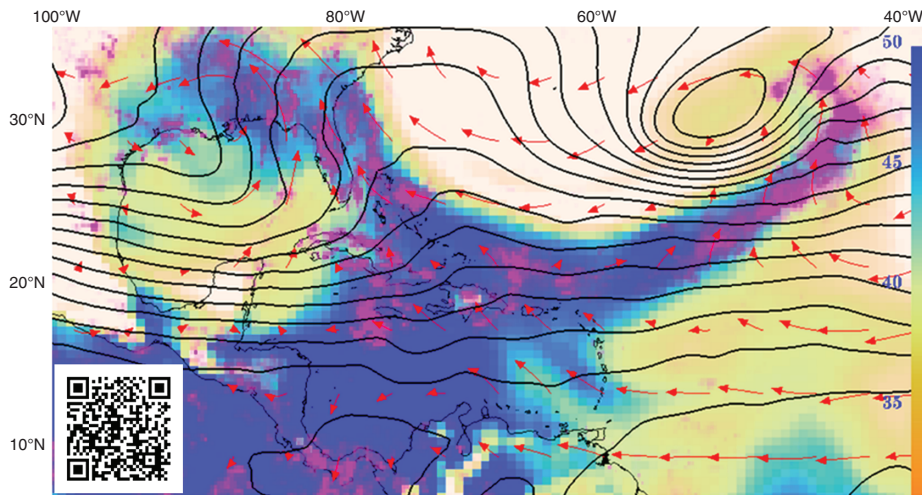
Figure 5. Synoptic map from 2009-05-23 showing daily column water vapour (mean of 1200 UTC and 0000 UTC in MERRA reanalysis, range 30–50 mm), daily accumulated rainfall from TRMM 3B42 (magenta shading, opaque for values >50 mm), and 1200 UTC flow at 850 hPa (curved red arrows are short-term trajectories) and 250 hPa (black contours are geopotential height at 20 m intervals). Scan the QR code with a smartphone to watch a YouTube video of this 2009 case study.

in the broader regional context of the Asian monsoon system with maps of monsoon onset date (Wang *et al.*, 2004; Wang, 2005). For a directly comparable depiction of Asian and North American monsoon onset progression, Figure 8 shows onset dates based on an objective wavelet analysis of the daily climatological time series of OLR (indicating deep convective clouds) at each point. Colour indicates the timing and strength of especially rapid decreases in OLR. Blue to red progressions indicating northward development over Asia are familiar from the many published maps of onset date, with annotations emphasizing the Indian Summer Monsoon, East Asian Summer Monsoon (EASM), and West Pacific Monsoon (WPM) which onsets rapidly starting in June. On the same scale and at the same latitudes and time of year as the EASM, we suggest that the Caribbean rain-belt may be viewed as an aspect of an analogous 'East North American Summer Monsoon' (ENASM).

Since the Caribbean rain-belt appears analogous to the EASM or Meiyu-Baiu in shape and seasonal evolution, perhaps its dynamics also are somewhat analogous. A recent study of Meiyu-Baiu dynamics by Sampe and Xie (2010, hereafter SX10) 'reveals the westerly jet as an important culprit for Meiyu-Baiu', both through its 'adiabatic induction of upward motion' by 'warm horizontal temperature advection' downwind of warm air of the southern Tibetan Plateau, and 'also by steering transient eddies … creating periods conducive to convection.' They further cite 'low-level southerly winds as another important environmental forcing for Meiyu-Baiu by supplying moisture'. We shall see that many of these aspects are also present over the Caribbean region, although we may refine interpretation of water vapour's effect on convection as occurring through its *presence*, not quite by its 'supply' inferred from mean winds acting as a 'forcing'.

Upper tropospheric westerly subtropical jets are common to all STCZs (Kodama, 1993). In contrast to winter fronts over land, subtropical jet-front systems have relatively weak surface temperature gradients. Instead, the thermal gradient that hydrostatically supports the height gradient of the upper level jet to satisfy thermal wind balance is located in the middle troposphere (Shapiro, 1981). Disruptions of balance by frontogenesis and/or by momentum advection cause secondary circulations, in this case in the upper troposphere. For instance, Molinari and Vollaro (2012) related deep clouds in the subtropical Northwest Pacific to 'the dynamical forcing of vertical motion' (Q vector divergence) in the 400–200 hPa layer. It remains somewhat unclear how such upper level quasi-geostrophic (QG) forcing is communicated to the lower troposphere where convection is governed.

Midlevel warm advection is another cause often cited for upward motion. In one view, warm advection is one of the two terms in the 'traditional' QG omega equation, as discussed in all synoptic textbooks (e.g. Lackmann, 2011). In a simpler view, SX10 emphasized its importance directly in the thermodynamic energy equation. When horizontal flow crosses T contours on an isobaric surface, yet those T contours are not moved (as is true of time mean flow), it implies a countervailing vertical advection by cross-isobaric (vertical) motion. More succinctly, a time-mean horizontal flow crossing a time-mean temperature gradient indicates adiabatic flow up a sloping isentropic surface (*isentropic upglide*). Such ascent can be further steepened in the tropics where ascending flow experiences diabatic heating from condensation and cloudy radiation. Can such an indication be inferred to be 'an external forcing', related to the warmth of air over the Tibetan Plateau, and 'inducing' the ascent, as in SX10's framing? Perhaps our geographical comparison can shed additional light on this question.
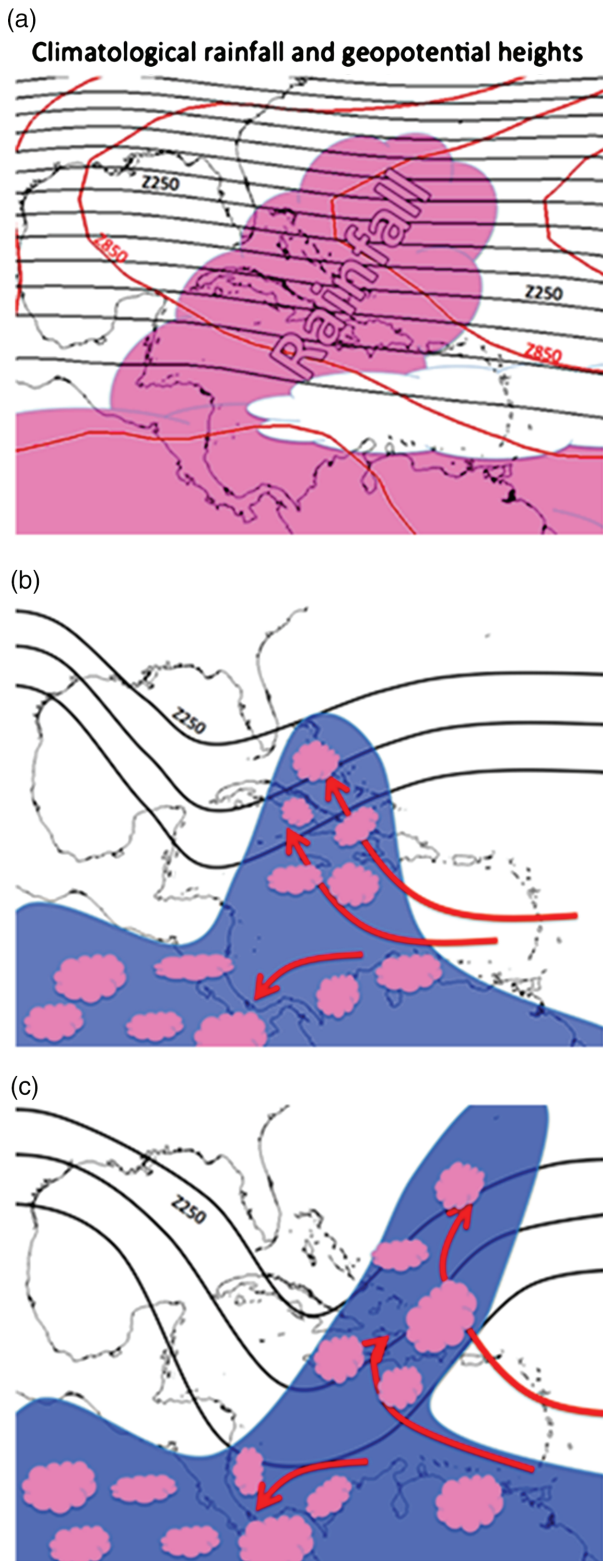
Figure 6. Schematic diagram representing the factors that contribute to the Caribbean rain-belt, based on imagery like Figure 5. Troughs in the westerlies induce southerly flow at low levels, which transports vapour northwards (blue tongues), especially if the trough is slow-moving or cuts off. Convective storms occur somewhat randomly within that moist air mass (pink), enhanced by quasi-geostrophic lifting ahead of the upper-level trough if it is strong. Diffluent flow associated with the Panama Low contributes a minimum in the western Caribbean Sea.

Figure 9 shows mean vertical motion and the advection of the time-mean T gradient by the time-mean flow in April, ERS, and July averages. Between the midlatitude westerlies and tropical easterlies, anticyclones straddle a temperature gradient. In the ERS in particular, the flow at 500 hPa (Figure 9(c)) features an anticyclone with southerlies and warm advection (red contours in Figure 9(d)) over the Caribbean rain-belt region, and its corresponding upward motion (shading in Figure 9(a), (c) and (e)). In Figure 9, the temperature gradient is smoothly meridional, and lacks the eye-catching warmth at its western end emphasized by the hot colours of Figure 6 of SX10. Rather than viewing the implications of the time-averaged temperature budget as a forcing by topographic influences unique to Asia, perhaps a more agnostic view is that mean ascent in STCZs might be diagnosed or indicated by such mean-flow advection of mean warmth at midlevels, *whatever its physical cause*.

Perhaps the more time-resolved interpretations of SX10 are more incisive. Their arguments emphasize the rectified (cumulative) rainfall that occurs in transient disturbances, noting that this can 'explain a paradox' when time-mean fields are inappropriately used to make climate arguments. Diagnostic relations in *averaged* fields can be subtly misleading about causality – and thereby could misdirect our search for sources of variability and possible climate influences on the system.

6. Summary and revisiting of the motivating question of climate modulation

To summarize, climatology (Figures 1–3) of accumulated rainfall during the ERS shows a SW-NE band of rainfall, here called the *Caribbean rain-belt*, which extends from the Caribbean coast of Central America towards The Bahamas and occurs on average between mid-May and mid-June. The Caribbean rain-belt is not a globally unique feature; similar seasonal rain-belt patterns are observed elsewhere at subtropical latitudes. The dynamics of the Caribbean rain-belt were compared to the Meiyu-Baiu, a well-studied (SX10 and references therein) subtropical rain-belt over the western Pacific region. Our inquiry led us to emphasize low-level flow, via its steering or shaping effect on the TPW field (the envelope for semi-random convective rain events) through northward displacement of humid air columns from the deep tropics. In this view, strong transient influences originating at upper levels are important, but perhaps indirectly, through their effects at low levels. Also, steady (if weak) low-level flow components, especially the meridional flows component the Panama Low, may also be especially important, since persistent flow can displace tracers like water vapour very effectively.

A major motivation for this study of this climatological regional climate system was that understanding its dynamics might point us to possibly predictable sources of variability, especially remote effects that might not be as obvious as, say, local SST. Inter-annual variability of
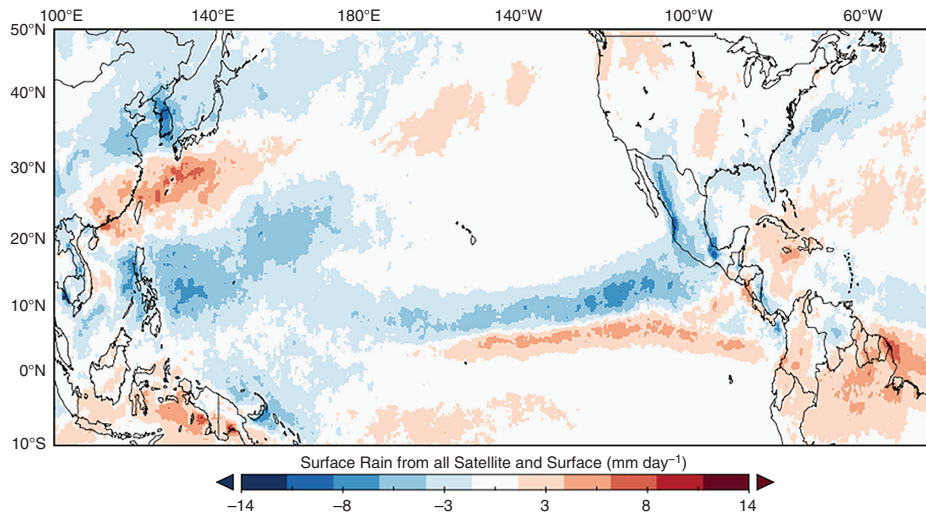
Figure 7. Climatological rain rate of the Early Rainfall Season, minus the July mean, from TRMM 3B42 data 1998–2012. Red colours indicate areas where ERS rain exceeds July rain: In the subtropics, this is over the Greater Antilles and in a belt just south of Japan.
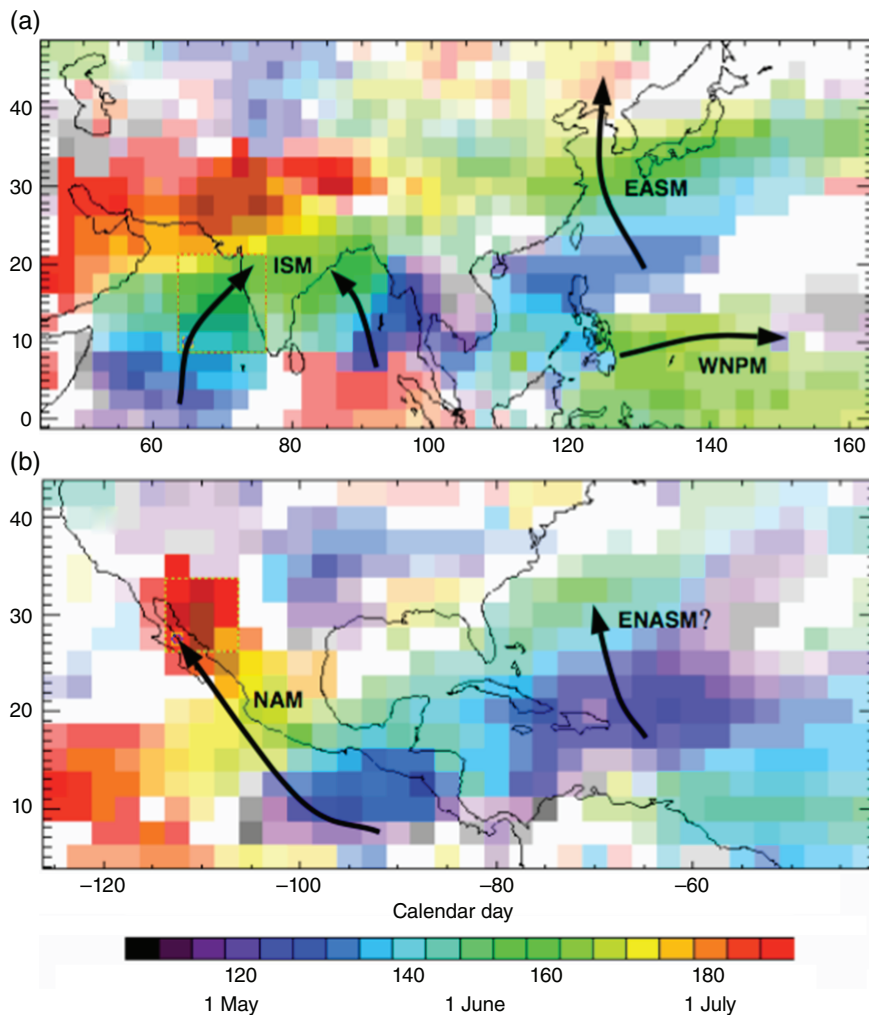

Figure 8. Date of sharp 'onset' of summer 'monsoon' cloudiness, derived from the climatological OLR time series at each point. The indicated date at each point is defined by the temporal location of the largest imaginary Paul wavelet power peak in the subseasonal band (25–140 day period) with its phase corresponding to a sharp decrease of the time series. The hue indicates date as indicated in the colour legend, while the density of coloration is proportional to the wavelet power peak's amplitude. If no power peak with a sufficient statistical significance index exists within the time interval, the grid cell is white, indicating there is no sharp increase in deep cloudiness in the spring–summer period indicated.
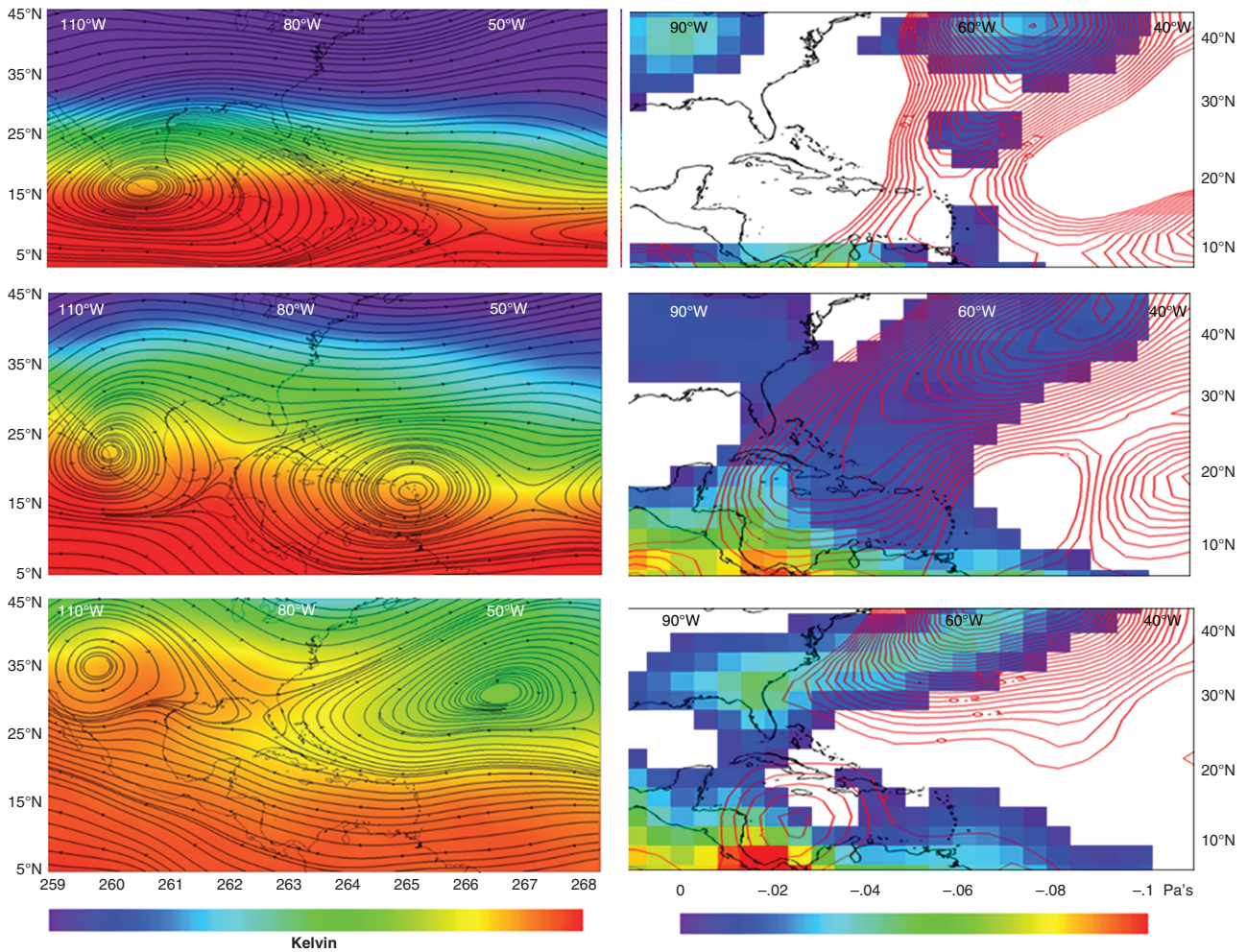
Figure 9. (a, c, e) Average 500 hPa temperature and 500 hPa streamlines for April (a), the Early Rainfall Season (c), and July (e). Average baseline period calculated from 1998 to 2012. (b, d, f) Average horizontal positive temperature advection (contours, 0.03 Ks-1 interval) and vertical velocity (shaded) at 500 hPa for April (b), the Early Rainfall Season (d), and July (f). Climatology calculated from 1998 to 2012. Average baseline period calculated from 1998 to 2012.

the Caribbean rain-belt impacts farmers who depend on the regular cycle of rainfall during the ERS. The first author's outreach activities and capacity building workshops throughout the Caribbean revealed a real demand for predictions of ERS rainfall in an effort to increase farmer resilience to inter-annual rainfall variability.

To revisit this question, it is useful to view some year-to-year variability. Figure 10 shows how 15 individual years relate to their climatological mean. Large year-to-year irregularity suggests that just a few disturbances or statistical degrees of freedom per ERS, supporting the idea of a synoptic control like the TPW tracer statistics. Meanwhile, very spotty spatial texture suggests the importance of convective and mesoscale processes in the actual rain making events, not highly forced storms like trough-driven fronts or squalls that would paint rainfall across the Earth's surface semi-uniformly.

Figure 10 shows that the Caribbean rain-belt was prominent during 2012 and 2009 (motivating the choice in Figure 5), but is hard to identify in other years (1998, 2001, and 2004). In some years (2008 and 2011), the

Caribbean rain-belt is more zonal with a very weak southwest to northeast slope. Our hypothesis that TPW tongues as essential would suggest that these aspects might be usefully related (in future research) to the *latitudinal shear at low levels*, and thereby to factors that influence that including possibly global momentum dynamics like those explored in Kelly and Mapes (2011, 2013, 2016).

If the main mechanism identified by SX10 – warm horizontal temperature advection at 500 hPa downwind of the Tibetan Plateau – were essential to Meiyu-Baiu rainfall, it might imply that Tibetan warmth (perhaps involving spring snowpack) would modulate Meiyu-Baiu rainfall year-by-year. But advection involves both a gradient and wind, so notwithstanding any anomalous mid-tropospheric warming, variability of a Eurasian westerly wave train embedded in westerly flow aloft remains a primary factor to the variability of the Meiyu-Baiu pattern (Kosaka *et al.*, 2011).

Low-level southwesterly flow along the western edge of the subtropical high also acts as a moisture source to feed the rain-belt. Might its mechanisms governing that flow,
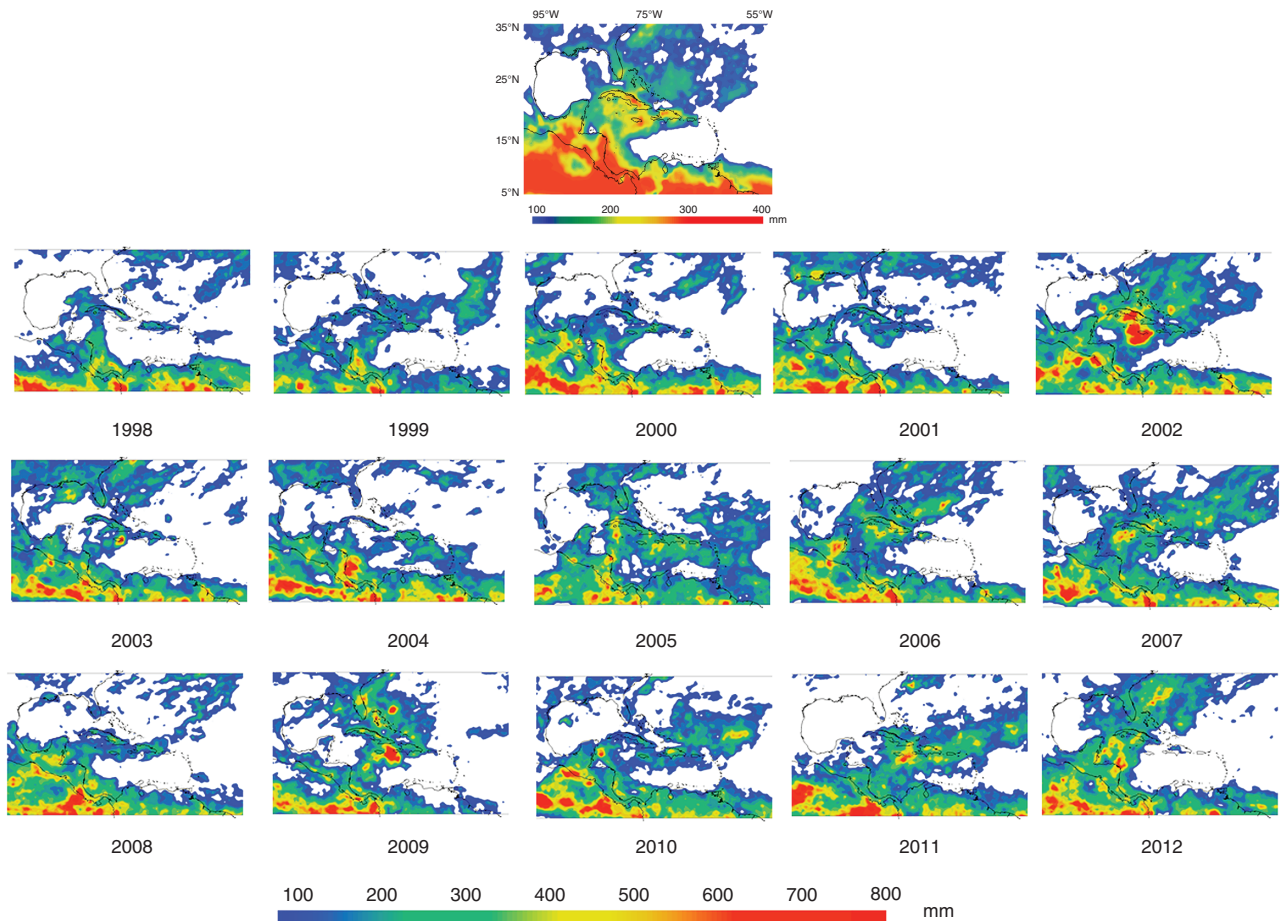
Figure 10. Accumulated Early Season rainfall for the 15-year mean (a) and the 15 individual years from 1998 to 2012 in TRMM 3B42 data (b). Rainfall accumulation under 100 mm is not shaded.

especially perhaps some weak but quasi-steady aspects of the flow, be an additional source of interannual variability? Subtropical highs are forced partly from the east as Rossby waves (Miyasaka and Nakamura, 2005).

Additional evidence for our hypothesis of TPW control of Caribbean rainfall is the weather observed during the drier winter months. Mid-latitude fronts during the winter can advance into the Caribbean (Liang, 2004), but rainfall along their path tends to be low. Winter flooding in the Caribbean is not totally uncommon, for instance the Christmas Day floods of 2012, which were associated with an out of season easterly wave. Usually, winter is dry because SST throughout the Caribbean Sea is well below convective instability thresholds. Southerly low-level flow out of the tropics is observed during the dry winter season, but energetic troughs and their low-level cyclones and fronts propagate through the Caribbean too quickly for their southerly flow and related moisture flux to persist as it does during the ERS. This illustrates the point that slow-moving transients have more effect, and suggests another set of variables that might modulate rainfall through the speed and scale (not just strength) of disturbances.

Rainfall does occur within the Caribbean during the MSD despite the absence of a westerly jet; however,

scattered rainfall during that time is largely driven by instability from local diabatic processes; note the island-locked rainfall pattern in Figure 3(c). Over land, rainfall during the mid-summer is less dependent on high-TPW tongues.

Distinguishing between forced ascent by upper-level PV anomalies (troughs) *versus* the shaping of the moisture tracer field by low-level winds induced by upper-level PV anomalies may help us test our hypothesis. Unfortunately, these effects co-occur, as our case study indicates. To achieve statistical significance, we need many cases, which means we must devise measures (algorithms) that convincingly discriminate between these correlated patterns. Such an exercise is beyond the scope of this study. Broadly speaking, such a discriminator might exploit the linearity of induced flow *versus* the nonlinearity (through advection terms) inherent in QG forcing for ascent. The two effects might therefore scale differently with amplitude (strength) of the upper-level feature, a property illustrated by the strong front near the intense trough *versus* the broader tropical rain belt in our case study.

Another outcome from this research may be (yet another) alternative explanation for the onset of the MSD; several mechanisms were mentioned in Section 3. But if northward shift of the transients in the westerlies during mid-summer is essential, then the *mechanism of impact on*

rainfall when easterlies replace westerlies (driven by Asian monsoon momentum dynamics; Kelly and Mapes, 2013, 2016) may need to be refined. Those authors emphasized the zonal displacement of the semi-permanent subtropical high and its weather conditions, but perhaps the steering of transients might be of first-order importance. Again, interannual variations might provide a test bed for distinguishing these mechanistic hypotheses, as well as a reason to care which one is more correct. To reiterate again, this research on characterizing a climatological rain pattern could for instance guide parameter selection for statistical modelling, motivated by applications that could benefit from forecasts.

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