1. Introduction

1.1 Motivation and objectives

The predecessor rain event (PRE) was first defined by Cote (2007, hereafter C07) as a coherent mesoscale region of heavy rainfall that develops as moisture originating from the vicinity of a tropical cyclone (TC) is transported by the synoptic-scale flow towards a region of atmospheric lifting well ahead of the TC. Sustained by deep tropical moisture characterized by precipitable water (PW) values often in excess of 50 mm, PREs are frequently manifested as persistent bands of heavy rainfall, with rainfall rates in excess of 100 mm (24 h)$^{-1}$. They therefore possess significant potential to cause extreme rainfall totals and high-impact flooding, an aspect that has been highlighted by the recent devastating flood-producing PREs associated with TC Frances (2004), TC Katrina (2005; C07), TC Erin (2007; Galarneau et al. 2010, hereafter GBS10), and TC Ike (2008).

Due to the substantial focus likely paid to local effects of landfalling TCs (i.e., high winds, storm surge, outer rainbands, eyewall convection) by operational forecasters and the lack of attention paid to their possible remote effects, PREs have the potential to take both the operational forecasting community and the general public by surprise. Adding to this problem, PREs may be poorly forecasted by numerical models due to their inability to: 1) adequately resolve mesoscale features and characteristics associated with heavy-rain-producing convective systems (Davis et al. 2003; Jankov and Gallus 2004), and 2) represent the dynamical and thermodynamic environmental influences of a TC moving into the midlatitudes (e.g., Atallah and Bosart 2003). Faced with these challenges, improved forecasting of PREs relies, in large part, on forecasters’ ability to recognize synoptic-scale patterns and key ingredients that could be potentially favorable.
for PRE development ahead of a TC. Specifically, we argue that a successful PRE forecast must involve a thorough identification of the processes that might lead to: 1) moisture transport from the TC to the PRE region, 2) sufficient forcing for upward vertical motion and thermodynamic instability to support heavy, convective rainfall in the presence of the TC moisture, and 3) quasi-stationary behavior of the precipitation system.

The goal of this thesis is to demonstrate, in a composite and case study framework, that there exist distinct, favorable pathways to PRE development, each representing a unique synoptic-scale configuration in which the three key factors: TC moisture, forcing for ascent, and slow precipitation system motion, culminate in the production of remote heavy rainfall ahead of a TC. Furthermore, we seek to elucidate the dynamical mechanisms that govern the spatial and temporal characteristics of PREs within these synoptic-scale configurations. Accomplishing these objectives will ultimately lead to the development of a set of operational forecasting tools and techniques not only to assess the potential for PRE development but also to project how it may evolve in time and space.

1.2 Background

1.2.1 Previous work on PREs

Though the PRE was only recently defined by C07, the concept has its origins in a study by Bosart and Carr (1978), which examined a heavy rain event in western New York and Pennsylvania ahead of TC Agnes (1972). Bosart and Carr documented how a stream of deep tropical moisture from the TC Agnes circulation interacted with a baroclinic environment, producing > 200 mm of rainfall well ahead of TC Agnes. The
remote region of heavy rainfall developed as moisture from TC Agnes streamed northward along the Appalachians towards a region of confluent flow and quasi-geostrophic (QG) forcing for ascent associated with an approaching midtropospheric short-wave trough. This process, illustrated schematically in Fig. 1.1, shows that the PRE developed remotely from the rain region directly associated with TC Agnes and, moreover, that it was dynamically driven by processes external to the TC.

The concepts expounded by Bosart and Carr (1978) on antecedent heavy rainfall ahead of TCs were recently extended by C07, who conducted a climatology of PREs associated with Atlantic basin TCs during 1998–2006. C07 identified 47 PREs associated with 21 TCs and stratified them by their position relative to the total observed track of the TC, finding that 26 occurred left of the TC track (LOT), 12 along track (AT) and 9 right of track (ROT). On average, PREs were separated from their parent TCs by approximately 1000 km and preceded the passage of the TC past the latitude at which the PRE occurred by ~36 h. Through composite analysis and case studies, C07 showed that PREs tended to form in a region where a poleward moisture surge from a TC interacted with a region of lift associated with a low-level baroclinic zone and the equatorward entrance region of an upper-tropospheric jet streak (Fig. 1.2a). In addition, C07 suggested that regions of orographic lifting and mesoscale boundaries associated with coastal fronts and regions of cold air damming could also be loci for PRE development (Fig. 1.2b). A recent study by Srock and Bosart (2009) showed that a coastal front on the oceanward side of a region of cold air damming in the lee of the southern Appalachian Mountains acted as the focus for antecedent heavy rainfall ahead of TC Marco (1990).

GBS10 examined the environments of PREs that occurred during 1995–2008 in a
storm-relative composite framework and through a case study of a high-impact PRE associated with TC Erin (2007). They documented the importance of the interaction of a region of deep baroclinicity beneath the equatorward entrance region of an upper-tropospheric jet streak with a poleward surge of moisture from the TC for PRE development. Specifically, frontogenesis and warm-air advection, driven by the impingement of strong low-level flow between the TC and an anticyclone to its east upon a low-level baroclinic zone, serves as a focus for PRE development, while further dynamical support is provided at upper levels as diabatically generated low potential vorticity (PV) outflow associated with the TC acts to strengthen the upper-tropospheric jet streak downstream of an advancing trough.

1.2.2. Relationship of PREs to heavy rainfall events

The results from the C07 and GBS10 studies suggest that the synoptic and mesoscale conditions associated with PREs closely resemble those associated with the “frontal” pattern for flash-flood-producing mesoscale convective systems (MCSs; Fig. 1.3) first documented by Maddox et al. (1979). The environmental properties and dynamical processes associated with this configuration are well documented (e.g., Maddox et al. 1979; Augustine and Caracena 1994; Glass et al. 1995; Junker et al. 1999; Moore et al. 2003; Schumacher and Johnson 2005, 2006) and typically are dominated by the advection of warm, moist air by a low-level jet (LLJ) towards a quasi-stationary low-level baroclinic zone (Figs. 1.3a,b) downstream of a weak 500-hPa short-wave trough (Fig. 1.3c). In this configuration, warm-air advection, moisture convergence, and convective destabilization maximized at the intersection of the LLJ with the baroclinic
zone provide the necessary ingredients to initiate and maintain MCSs. Additionally, frontogenesis associated with horizontal speed convergence and horizontal deformation at the terminus of the LLJ has been shown to be an important mesoscale lifting mechanism to support MCSs (e.g., Trier and Parsons 1993; Augustine and Caracena 1994; Trier et al. 2006).

Within an environment of sufficient moisture, instability, and lift to support convection, prolonged heavy rainfall accompanying PREs and MCSs associated with the “frontal” pattern is favored by slow system movement and convective cell “training” (e.g., Doswell et al. 1996), a process by which convective cells repeatedly pass over a given area. Most frequently the mesoscale organization of PREs and heavy-rain-producing MCSs resembles the “training line/adjoining stratiform” (TL/AS) and “backbuilding/quasi-stationary” (BB) extreme-rain-producing MCS archetypes (Fig. 1.4) defined by Schumacher and Johnson (2005). Both of these organizational modes characterized the PRE associated with TC Erin (2007), examined by GBS10. In this event, a line of convection oriented parallel to the 0–6 km vertical wind shear resulted in the repeated passage of convective cells over southern Minnesota and western Wisconsin (TL/AS organization), while new cells continuously developed directly upstream (BB organization), keeping the line stationary. Such a combination of the TL/AS and BB organizational modes allows for high rainfall rates to persist for a prolonged period of time, therefore posing a substantial risk for locally extreme rainfall totals and flash flooding (e.g., Chappell 1986; Doswell 1996; Junker et al. 1999; Schumacher and Johnson 2005).

For TL/AS MCSs (Fig. 1.4a), veering wind direction with height at low levels and
weak-to-moderate midlevel speed shear oriented parallel to the axis of convection favor slow, line-parallel MCS motion (e.g., Maddox et al. 1979; Junker et al. 1999; Schumacher and Johnson 2005). For BB MCSs (Fig.1.4b), persistent moisture convergence and mesoscale lifting, afforded by the interaction of a low-level jet with a surface frontal boundary, favors the continuous development of convective cells upstream of an antecedent convective line (e.g., Chappell 1986; Glass et al. 1995; Doswell et al. 1996; Schumacher and Johnson 2005). In addition, environments characterized by high relative humidity and, correspondingly, high PW values, which are typical of PREs (GBS10), favor high precipitation efficiencies (Market et al. 2003) and preclude the formation of strong downdrafts, which would otherwise favor forward system propagation (e.g., Doswell et al. 1996; Davis 2001).

On the synoptic scale, dynamical support for heavy rainfall associated with MCSs and PREs is often provided by an upper-tropospheric jet streak (e.g., Uccellini and Johnson 1979) or by an approaching short-wave trough (e.g., Bosart and Carr 1978; Doswell and Bosart 2001). Though typically not sufficient to trigger convection, the QG forcing for ascent associated with an ageostrophic circulation accompanying a jet streak or associated with upward increasing cyclonic vorticity advection linked to a short-wave trough can aid in both environmental moistening and thermodynamic destabilization, thereby making conditions favorable for convection (Doswell 1987).

The indirect and direct thermal circulations associated with an upper-tropospheric jet streak can couple with the thermally direct circulation associated with a region of lower-tropospheric frontogenesis, thus producing troposphere-deep regions of ascent (Keyser 1999; Hakim and Keyser 2001). Typically, the superposition of lower- and
upper-tropospheric thermally direct circulations, illustrated in Fig. 1.5, occurs frequently in environments of heavy-rain-producing MCSs in the warm season (e.g., Junker et al. 1999; Moore et al. 2003) and in the environments of PREs (C07; GBS10). As these thermally direct circulations couple, they can be further augmented by latent heat release within a persistent region of heavy rainfall. This heating can act to strengthen thermal gradients in the middle troposphere and thereby enhance frontogenesis (Hsie et al. 1984; Colle 2003). In addition, the diabatically generated anticyclonic outflow at upper levels associated with the region of heavy rainfall can strengthen the upper-tropospheric jet streak, leading to an enhanced thermally direct circulation in the entrance region (e.g., Maddox et al. 1981; Keyser and Johnson 1984; Wolf and Johnson 1995).

In some PRE cases, such as TC Fran (1996) studied by C07 in which baroclinicity and QG forcing for ascent were weak, persistent moist upslope flow along orographic features, such as the Appalachian Mountains, can be the main focus for the development of heavy rainfall. Elevated terrain can serve as an anchor for quasi-stationary heavy rainfall provided that the flow remains upslope and that moisture and instability are continuously replenished (e.g., Maddox et al. 1978; Lin et al. 2001). The lifting of moist, unstable air over elevated terrain has been frequently implicated as a mechanism for extreme rainfall in the western Mediterranean (e.g., Doswell et al. 1998; Romero et al. 2000) and along the northwestern coast of the U.S. (e.g., Ralph et al. 2005), and can also result in the enhancement of precipitation directly associated with landfalling TCs (e.g., Schwarz 1970; Sinclair 1994).
1.2.3 PREs in the context of extratropical transition of TCs

Though the environments and the physical mechanisms associated with PREs closely resemble those classically implicated for flash flooding and heavy rainfall, PREs should, nonetheless, be treated separately from ordinary heavy rain events due to both the direct and indirect dynamical and thermodynamic influences of a TC on their formation. Indeed, analogies can be drawn between the PRE process and the extratropical transition (ET) process, though we stress that they are two distinct processes. The composite analyses and case studies conducted by C07 and GBS10 indicate that the Maddox et al. (1979) “frontal” pattern in the environments of PREs often arises as strong poleward low-level flow associated with the outer circulation of the TC, aided by anticyclonic flow to its east, impinges upon a low-level baroclinic zone beneath the equatorward entrance region of an upper-tropospheric jet streak. This setup, characterized by veering winds with height at low levels in the presence of warm-air advection, frontogenesis, and deep moisture flux from the TC, favors the formation of a quasi-stationary region of heavy rainfall along the baroclinic zone. This configuration is analogous to the transformation stage of ET (Fig. 1.6), documented by Klein et al. (2000), wherein frontogenesis and warm-air advection serve as lifting mechanisms for the development of heavy rainfall ahead of a poleward-moving TC as its circulation begins interacting with a pre-existing baroclinic zone. The key distinction for PREs in this context is that they are separate entities from the main TC rain shield.

Additionally, as is common for PREs (GBS10), the ET process involves the favorable interaction of a TC with an approaching middle-/upper-tropospheric trough (e.g., DiMego and Bosart 1982a,b; Harr et al. 2000; Atallah and Bosart 2003). During
this interaction, the development of large vorticity gradients in the middle and upper troposphere between the approaching trough and the diabatically generated outflow ridge associated with the TC leads to enhanced QG forcing for ascent associated with cyclonic vorticity advection by the thermal wind (e.g., Bosart and Lackmann 1995; Klein et al. 2002; Atallah and Bosart 2003; Atallah et al. 2007). This forcing for ascent, in turn, favors the formation of asymmetries in the TC rain shield and contributes to the reintensification of the TC as an extratropical cyclone.

As the TC undergoes ET, the reduction of PV and attendant ridge building in the upper troposphere due to diabatic processes (i.e., latent heating) associated with the TC rain shield can result in enhanced downstream upper-tropospheric PV gradients (or similarly geopotential height gradients), leading to the strengthening of a downstream upper-tropospheric jet streak (e.g., Bosart and Lackmann 1995; Atallah and Bosart 2003; Agustí-Panareda et al. 2004). Reimer et al. (2008) showed the intensification of a downstream upper-tropospheric jet streak occurs during ET in association with the poleward advection of low-PV air by the diabatically generated divergent outflow associated with a transitioning TC. As a jet streak strengthens in response to diabatic processes, the ageostrophic circulation within the entrance region can be enhanced, which leads to greater ascent and enhanced heavy rainfall poleward of the TC center (e.g., Bosart and Lackmann 1995; Atallah and Bosart 2003). The composite analysis conducted by GBS10 suggests that the strengthening of an upper-tropospheric jet streak in response to diabatic heating and attendant ridge amplification accompanying the TC is likely an important process in PRE development.
1.2.4 Remote rainfall associated with TCs in the North Pacific

Recent studies have investigated the influence of TCs in the Western and Eastern North Pacific on distant rainfall, suggesting that conditions favorable for PREs are not limited to Atlantic basin TCs. Wang et al. (2009) showed how moisture advected by the outer circulation of Typhoon Songda (2004) contributed to distant heavy rainfall in Japan as it recurred along the eastern coast of Asia. Eastern North Pacific TCs have been linked to moisture surges in the Gulf of California, which in turn impact rainfall in northwestern Mexico and the southwestern United States (e.g., Higgins and Shi 2005; Farfán and Fogel 2007). Recent work by Corbosiero et al. (2009) and Svoma (2010) has documented the important contribution of eastern North Pacific TC moisture to rainfall distributions over the southwestern United States.

1.3 Thesis format

The remainder of this thesis is organized as follows. Chapter 2 will document the various data sources and methods employed to identify and stratify PRE cases, to construct composite analyses, and to conduct case studies. A climatology of PREs that occurred during 1988–2008 will presented in chapter 3, and PRE-relative composite analyses will be presented in chapter 4. Chapter 5 will present multiscale case studies of PREs associated with TC Rita (2005), TC Wilma (2005), and TC Ernesto (2006). The results of this thesis will be summarized and discussed in chapter 6. Chapter 7 will provide concluding remarks and suggestions for future work.
Figure 1.1: Schematic midtropospheric streamlines associated with a PRE ahead of TC Agnes (1972). Shaded regions indicate loci of heavy rainfall. Reproduced from Fig. 13 in Bosart and Carr (1978).

Figure 1.2: Conceptual model of the synoptic-scale environment associated with LOT PREs in advance of TCs, revised and updated from Bosart and Carr (1978). Position of TC is given by tropical storm symbol. Representative TC tracks are marked with solid blue arrows. Low-level (LL) features are representative of the 925-hPa level, midlevel (ML) features are representative of the 700-hPa level, and upper-level (UL) features are representative of the 200-hPa level. Boxed region indicates the area of the mesoscale and physiographic conceptual model shown in panel (b). Reproduced from Figs. 5.1 and 5.2.
Figure 1.3: Schematic depictions of the (a) surface, (b) 850 hPa, and (c) 500 hPa patterns associated with the “frontal” type flash flooding scenario. The shaded box indicates the region of greatest potential for heavy rainfall and flash flooding. Figure reproduced from Figs. 8a–c in Maddox et al. (1979).
Figure 1.4: Schematic diagram of the radar-observed features of the (a) TL/AS and (b) BB patterns of extreme-rain-producing MCSs. Contours (and shading) represent approximate radar reflectivity values of 20, 40, and 50 dBZ. In (a), the low-level and midlevel shear arrows refer to the shear in the surface-to-925-hPa and 925–500-hPa layers, respectively, as discussed in section 4 of Schumacher and Johnson (2005). The dash–dot line in (b) represents an outflow boundary; such boundaries were observed in many of the BB MCS cases. The length scale at the bottom is approximate and can vary substantially, especially for BB systems, depending on the number of mature convective cells present at a given time. Caption and figure reproduced from Fig. 3 in Schumacher and Johnson (2005).
Figure 1.5: Schematic cross section of an elevated convective event taken parallel to a low-level jet (LLJ) across a surface frontal zone. Dashed lines represent typical $\theta_e$ values, the large stippled arrow represents the ascending LLJ, the thin solid oval with arrows represents the ageostrophic direct thermal circulation (DTC) associated with the upper-level jet (ULJ), and the dash–dotted oval with arrows represents the DTC associated with the low-level frontogenetical forcing. The area aloft enclosed by dotted lines indicates upper-level divergence; the area aloft enclosed by solid lines denotes the location of the ULJ. Note that in this cross section, the horizontal distance between the MCS and the location of the ULJ is not to scale. Caption and figure reproduced from Fig. 14 in Moore et al. (2003).
Figure 1.6: Conceptual model of transformation stage of ET in the western North Pacific, with labeled areas as follows: 1) environmental equatorward flow of cooler, drier air (with corresponding open cell cumulus); 2) decreased TC convection in the western quadrant (with corresponding dry slot) in step 1, which extends throughout the southern quadrant in steps 2 and 3; 3) environmental poleward flow of warm, moist air is ingested into TC circulation, which maintains convection in the eastern quadrant and results in an asymmetric distribution of clouds and precipitation in steps 1 and 2; steps 2 and 3 also feature a southerly jet that ascends tilted isentropic surfaces; 4) ascent of warm, moist inflow over tilted isentropic surfaces associated with baroclinic zone (dashed line) in middle and lower panels; 5) ascent (undercut by dry-adiabatic descent) that produces cloudbands wrapping westward and equatorward around the storm center; dry-adiabatic descent occurs close enough to the circulation center to produce erosion of eyewall convection in step 3; 6) cirrus shield with a sharp cloud edge if confluent with polar jet. Caption and figure reproduced from Fig. 5 in Klein et al. (2000).