12.811 Term Paper — Exploring the Role of Surface Fluxes in the Reintensification of Tropical Storm Erin (2007)

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Abstract

The intensification of Tropical Cyclones (TCs) remains relatively poorly understood – detailed investigation of unusual cases of TC intensification may shed light on the process. Here, we investigate the role of surface enthalpy fluxes in the overland reintensification of TC Erin (2007). Calculation of the surface enthalpy flux by multiple methods using data from the Oklahoma Mesonet suggests that these fluxes were relatively small, $\sim$ 100 W m^{-2}, during reintensification. This is quite a bit smaller than suggested by Emanuel (2008), though similar in magnitude to the findings of Evans et al. (2010) and Evans et al. (2011) for simulated storms that reintensify. Straightforward use of potential-intensity theory indicates that fluxes of this magnitude are consistent with the observed strength of Erin. Use of a modified version of the CHIPS model shows that in an axisymmetric framework, even such weak surface fluxes can reintensify a nearly saturated warm-core vortex at a rate and to a strength consistent with observations of Erin. However, ambient conditional instability and advection of moist static energy into the core of the storm from the southeast are likely non-negligible sources of energy in comparison to surface enthalpy fluxes. The main challenges for the surface flux theory of reintensification of TC Erin are thus the importance of other energy sources, the dynamics of symmetrization, and the question of why overland reintensification is so uncommon.

1 Introduction

The reintensification of Erin has been described in detail by Arndt et al. (2009) and many of the Erin-specific references in this paper. Briefly, between 00 and 12 UTC, on Sunday the 19th of August, 2007, the remnants of the weak Tropical Storm Erin reorganized dramatically over southwest Oklahoma. The storm's central pressure dropped by about 12 hPa to 995 hPa, sustained 10-meter winds of 25 m s^{-1} were recorded (Knabb, 2008), and a band of strongly precipitating deep convection formed an eye-like structure on radar images (Figure 1) (MMM, 2011). This reintensification occurred over the Oklahoma Mesonet, a dense meteorological observation network with over 110 stations across the state, allowing for detailed analysis of the factors that contributed to the reintensification of the cyclone.

The reintensification of Erin has been the subject of some debate in the literature, a significant part of it based on whether terminology pertinent to "tropical" cyclones is even appropriate to the situation. Knabb (2008) defends the choice of the National Hurricane Center (NHC) to classify Erin as simply a "low" rather than a TC during reintensification over Oklahoma; the NHC "best-track" for Erin from this paper is shown below (Figure 2). Monteverdi and Edwards (2010) suggest that Erin was really an inland TC, and that the NHC misclassified the storm as non-tropical merely because it was over land and not over the ocean. Erin reintensified as a "warm-core non-frontal synoptic-scale cyclone", "with organized deep convection and a closed surface circulation about a well-defined center," in a region with very weak baroclinicity, and thus meets all of the NHC criteria for TC classification except for heat extraction from an oceanic source. However, this latter condition is a weak semantic distinction, since the atmosphere does not care whether the source of enthalpy is ocean or land; thus we will take the view that Erin reintensified as a TC over Oklahoma.

As the vast majority of TCs weaken upon landfall, and this is the only known instance where a TC attained a lower surface pressure and stronger surface winds over land than at any time of its existence over the ocean, several studies have speculated about what unique factors could have led to the reintensification of Erin. Arndt et al. (2009) suggest that a weak upper-level shortwave trough provided a lifting mechanism to the east of Erin's remnants, and that the combination of a lifting mechanism and conditional instability allowed moist convection to reintensify the remnants of Erin. Evans et al. (2010) and Evans et al. (2011) continue this line of argument, and suggest

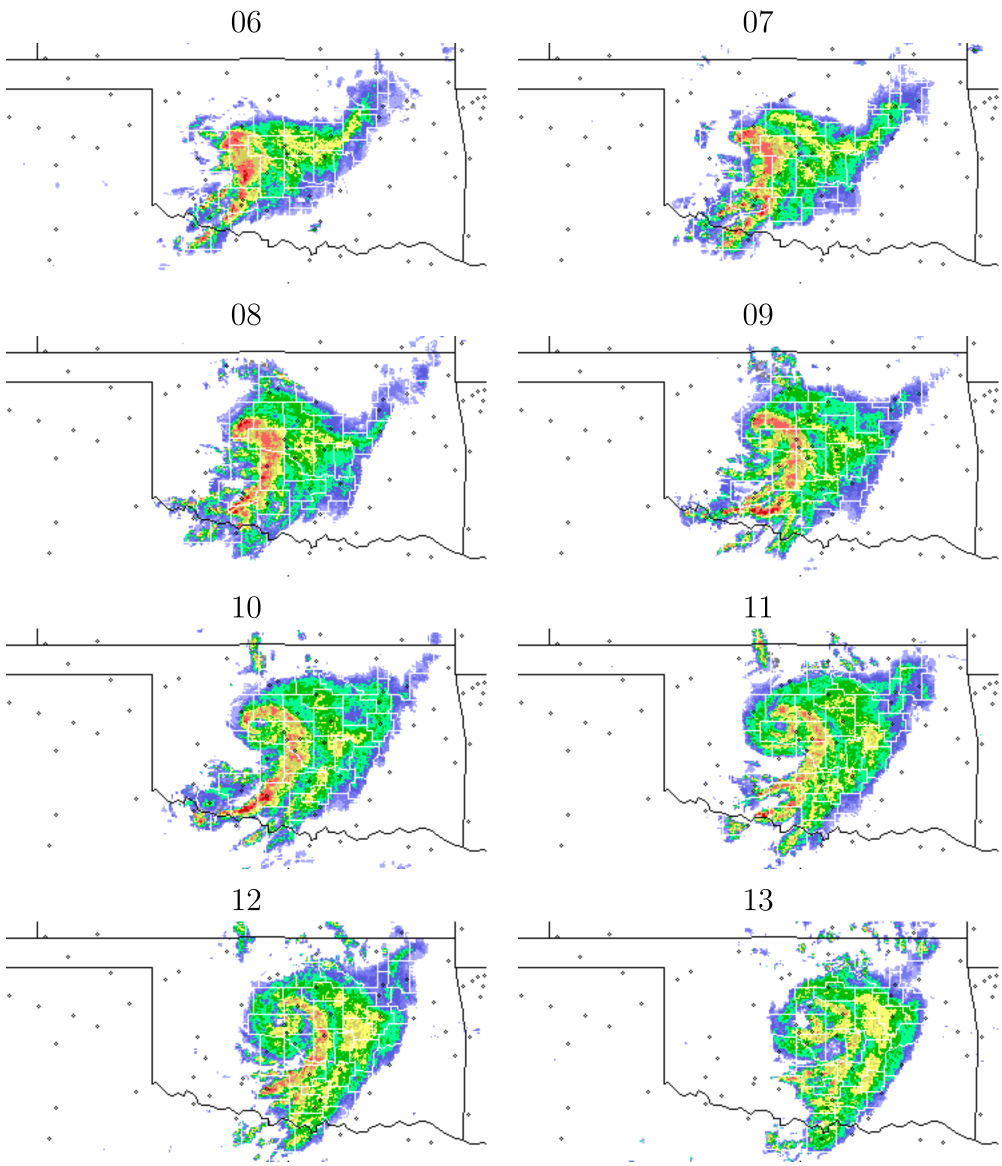

Figure 1: Radar images of reintensification of Erin on 19 August, 2007; UTC time in hours indicated above each subfigure. Constructed from images from MMM (2011)

that anomalously high soil moisture surrounding Erin is key to establishing the conditionally unstable environment which then serves as a source of

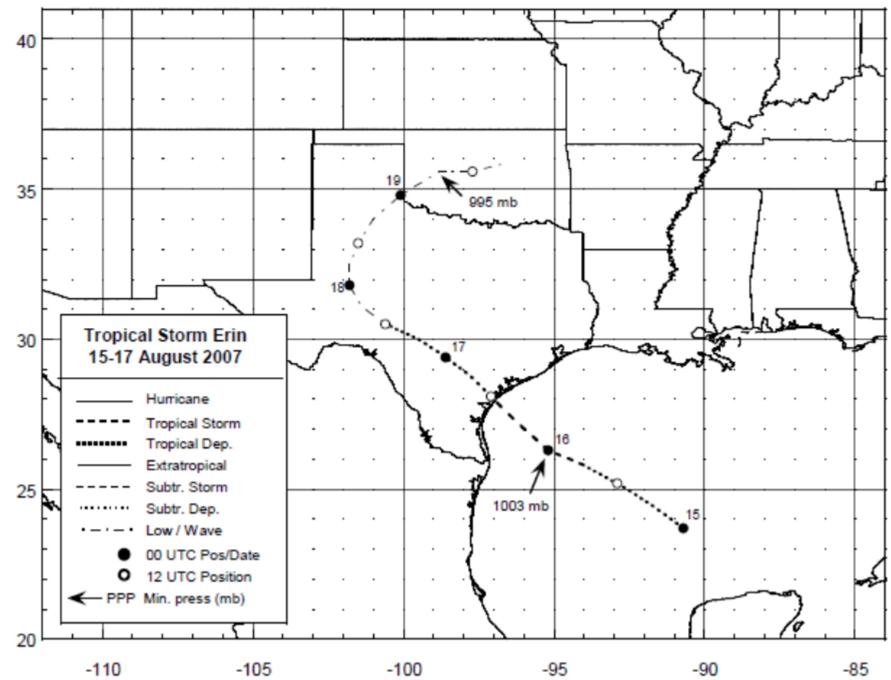

Figure 2: NHC best-track for TC Erin and its remnants – Knabb (2008) Figure 1.

available potential energy (APE) for Erin.

These arguments are correct in a sense, in that convection in an unstable environment can convert APE to kinetic energy of an organized, large-scale flow. However, Evans et al. (2010) and especially Arndt et al. (2009) fail to compare surface fluxes and conditional instability on an even footing. Also, CAPE for undiluted ascent of surface parcels is likely much larger than the path-integrated buoyancy for parcels in real (entraining) cumulus convection.

Normally, TCs accomplish surface heating locally, by exploiting the thermodynamic disequilibrium between the surface and the boundary-layer atmosphere (Emanuel, 1986). The oceanic mixed layer provides a large heat reservoir from which turbulent surface fluxes transfer enthalpy to the boundary-layer in most TCs, but there is no good *a priori* reason to think that an ocean surface must be the surface that supplies enthalpy to a TC, nor that the enthalpy flux must be predominantly latent heat. Misconceptions abound in the meteorological community about the crucial roles of absolute sea surface temperature, and of the necessity of latent heat fluxes at the surface, but there is nothing intrinsic about either. A recent study has shown that a dry tropical cyclone is quite possible (Mrowiec et al., 2011). In the broader context of TCs as warm-core, flux-driven vortices, Emanuel et al. (2008) hypothesize that when warm, dry soils are moistened by the leading rains of a TC, the increase in soil thermal conductivity can result in a dramatic increase in the surface heat flux and lead to intensification (or more likely reintensification) of a TC. This mechanism is proposed with regards to reintensification of TCs over the deserts of northwest Australia, where dark, mostly bare, sandy soils heat up over a deep layer, and have a much higher thermal conductivity when wet than when dry. Emanuel (2008), following the work of Emanuel et al. (2008) suggests that precipitation-induced conductive surface heat exchange (PICSHE?) may have driven the reintensification of Erin.

While the terrain of Oklahoma is quite different than northwest Australia, large swaths of sandy soils exist, there is a belt of winter wheat cultivation that is typically fairly bare after the early summer harvest, and the 2-week period prior to the passage of Erin was fairly hot, dry, and clear, allowing soils to store a significant amount of heat. However, the early summer in 2007 was anomalously wet over the southern plains, and despite the preceding two weeks of hot, dry weather, soil temperatures in early August 2007 in Texas and Oklahoma were likely below their climatological mean values. Using high resolution WRF simulations, Evans et al. (2010) and Evans et al. (2011) attempt to isolate the factors that allowed for the dramatic strengthening of Erin on the 19th of August. They hypothesize that the combination of an anomalously moist soil from early-summer rainfall, and rain in Texas from the initial passage of the storm, allow the formation of a shallow, moist, high-entropy boundary layer in the storm's "outer" environment, which is more important than "near-vortex" surface fluxes in allowing Erin to reintensify, but they do not compare near-vortex surface fluxes and broader environmental modification on energetic terms. The goal of this paper is to use data from the dense observational network of the Oklahoma Mesonet (McPherson et al., 2007, Brock et al., 1995) to assess the role of surface fluxes, relative to other factors, in the reintensification of Erin.

2 Inference of Surface Heat Flux

Soil temperature measurements are collected at most of the Mesonet observation sites, and are thus probably the best way to infer surface enthalpy fluxes across the domain of interest. Soil temperature is measured at 5-, 10-, and 30-cm depths below native sod, and at 5- and 10-cm depths below bare soil. Soil moisture is measured indirectly, by the heat-pulse method, at depths of 5-,25-,60-, and 75-cm at a number of the sites as well. All data were obtained from the internet, at addresses such as: http://www.mesonet.org/index.php/dataMdfMts/ dataController/getFile/*yyyymmddhhnn*/mdf/TEXT/, (no space) where *yyyymmddhhnn* is a time identifier (year-month-day-hour-minute).

We will examine two methods of quantifying the surface heat flux from soil temperature measurements. The first is conductivity-based; we can infer the vertical heat flux at the top of the soil by using measurements of the vertical temperature gradient and a model of the soil thermal conductivity. The second approach is energy-based; we can infer the surface heat flux as a residual of the energy balance of the top half-meter or so of soil. There are advantages and limitations to both these approaches, which we will discuss later, but an important caveat to both that should be mentioned up-front is that soils are highly heterogeneous, so even with a fairly dense observational network, there is no guarantee that important microscale features will be captured, and there is a great deal of variability in the strength of surface fluxes.

The upward conductive soil heat flux F_c is straightforward to model in a finite-difference framework:

$$F_c = k_T \frac{T_{z1} - T_{z2}}{z1 - z2}, \tag{1}$$

where k_T is the soil thermal conductivity in W m^{-1} K^{-1}, T_{z1} is the soil temperature at depth $z1$, and T_{z2} is the soil temperature at depth $z2$. The principal limitations to this approach are uncertainties in the value of the soil thermal conductivity, and inability to compute the soil heat flux at the surface directly from measurements, since we do not have the surface skin temperature. This latter issue is a problem only if a large fraction of the soil heat flux divergence occurs in the uppermost 5-10 cm, since we can estimate the soil heat flux at 5 cm based on soil temperature measurements at 5 and 10 cm depths. Typical values of soil thermal conductivity range between 0.5-2 W m^{-1} K^{-1} (Ochsener et al., 2001), but values can be higher for saturated sandy soils, or lower for very dry clay soils (Becker and Fricke, 1997). Free convection of water even in saturated sandy soil with a deep unstable layer is unlikely to significantly enhance the upward conductive heat flux, as the pertinent Rayleigh number is subcritical ($\sim 20 < Ra_c \sim 40$) even for an unrealistically pervious soil ($k \sim 10^{-10}$ m^2) with a large temperature gradient of 5 K over a half-meter deep layer (see Elder (1967) and Johansen (1975) for discussion of free convection in porous media).

Many models of soil thermal conductivity exist at different levels of complexity, in terms of mechanistic basis, empirical validation, and required input parameters (e.g. Becker and Fricke (1997)). Soil thermal conductivity depends strongly on soil water content and soil texture, though much of this dependence can be condensed into a dependence on volume fraction of air in the soil (Ochsener et al., 2001). In the interest of simplicity, we will only crudely explore the dependence of k_T on soil water. Illston et al. (2008) describe the soil moisture monitoring method of the Oklahoma Mesonet: *in situ* probes measure the rate of dissipation of a heat pulse within a porous ceramic matrix having similar water-adsorbing properties to a silt loam soil. The temperature change before and after passage of a heat pulse is indicative of soil moisture: lower soil moisture means that a probe dissipates less heat and warms quite a bit; higher soil moisture means that a probe dissipates more heat and warms less. The Fractional Water Index (FWI) is the Mesonet supported soil moisture variable, is linear in the probe temperature change, and is defined so as to be 1 for a saturated soil and 0 for a very dry soil. The FWI is not a direct indicator of soil moisture in terms of volumetric water content (the moisture variable conventionally used in models of soil thermal conductivity), but as it directly measures the rate of heat dissipation in a porous medium with similar adsorption properties as compared to soil, we believe it is a reasonable measured value to use for linearly scaling soil thermal conductivity between characteristic wet and dry values of $k_{T,w} = 2$ W m^{-1} K^{-1} and $k_{T,d} = 0.5$ W m^{-1} K^{-1}:

$$k_T = k_{T,d} \times (1 - \mathrm{FWI}) + k_{T,w} \times \mathrm{FWI} \tag{2}$$

Using equation 2, together with time-varying measurements of FWI and the 5-10 cm soil temperature gradient, we estimate that soil heat fluxes are only the order of tens of watts per square meter over the main region of redevelopment of TC Erin, when averaged over the time period of redevelopment (Figure 3). The location of maximum soil heat flux is mostly near the track of the cyclone, but slightly to the southeast (see Figure 2 for track). The heat fluxes as estimated by the conductive method are much lower than a single site estimate of 500 W m^{-2} by Emanuel (2008), and somewhat lower even than the 100-125 W m^{-2} estimated by Evans et al. (2010) and Evans et al. (2011). As noted above, this estimate fails to capture divergence of the soil heat flux in the surface 5-10 cm, and thus may be a substantial underestimate of the true surface heat flux. However, the small magnitude of the conductive heat flux at 5-10 cm acts as a substantial constraint on the surface heat flux, because we can rule out extremely large sustained energy loss by the top 5-10 cm of soil in any one location (a saturated soil column 7.5 cm deep cooling at 2 K per hour is losing approximately 125 W m^{-2}; such large surface cooling rates are rarely sustained for more than an hour in observations).

At its simplest, the energy-balance method just looks at the change in heat storage by dry soil. Thus, in a finite-difference approach, the energy-balance based surface flux, F_{eb}, is equal to the opposite of the time rates of change of the heat contents of a sum of measured soil layers

$$F_{eb} = -\sum_{i=1}^{n} \frac{\Delta(c_i h_i T_i)}{\Delta t}, \tag{3}$$

where the ith soil layer has volumetric heat capacity c_i, thickness h_i, and temperature T_i. Layer depths for the 5, 10, and 30-cm soil temperature measurements are considered to be 7.5, 12.5, and 30 cm, respectively, and cumulatively span the top half-meter of soil. Bulk soil density is measured at each site, and used together with information about total porosity (equal to site-specific saturated water content) to infer the density of soil solids. Soil solid density, together with a soil-solid heat capacity of 850 J kg^{-1} K^{-1}, and site-specific residual water content, are used to calculate a dry-soil volumetric heat capacity. Across all sites, dry soil volumetric heat capacity ranges from roughly 1-2 $\times 10^6$ J m^{-3} K^{-1}, and saturated soil volumetric heat capacity ranges from roughly 2.7-3.3 $\times 10^6$ J m^{-3} K^{-1}. Use of a constant dry soil heat capacity in equation 3 gives maximum sustained surface fluxes on the order of 50 W m^{-2} (Figure 4).

The major problems with the dry-soil energy-balance method are that changes in soil water content over the period of consideration due to precipitation alter the soil heat capacity significantly, incoming rainwater is colder than the soil, and if there is percolation of water out of the bottom boundary of our half-meter soil column, the system is not energetically closed. A

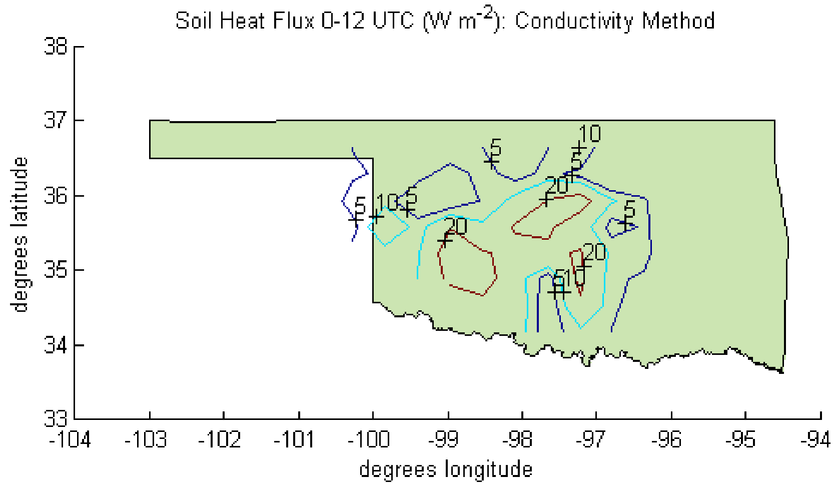

Figure 3: Estimate of time-mean upward soil heat flux at a depth of 5-10 cm, from 0-12 UTC on August 19, based on equations 1 and 2. Soil temperature data from roughly 80 of the Mesonet sites are interpolated to a 20 by 10 grid and the conductive heat flux is calculated on the grid each hour from 0-12 UTC, then averaged.

decrease in soil temperature accompanied by an increase in soil water content thus may reflect an increase, decrease, or no change at all in the heat content of the rainwater+dry soil system. Expanding equation 3 in terms of differentials of $\Delta c_i/\Delta t$ and $\Delta T_i/\Delta t$ gives:

$$F_{eb} = -\sum_{i=1}^{n} h_i T_i^* \frac{\Delta c_i}{\Delta t} + h_i c_i \frac{\Delta T_i}{\Delta t}. \tag{4}$$

Here, the temperature used in multiplying the time rate of change of the volumetric heat capacity (T_i^*) is the change in temperature experienced by the added heat capacity, or $T_i - T_{\text{rain}}$. If we assume that changes in FWI reflect changes in the volumetric soil moisture (vsm) relative to the difference

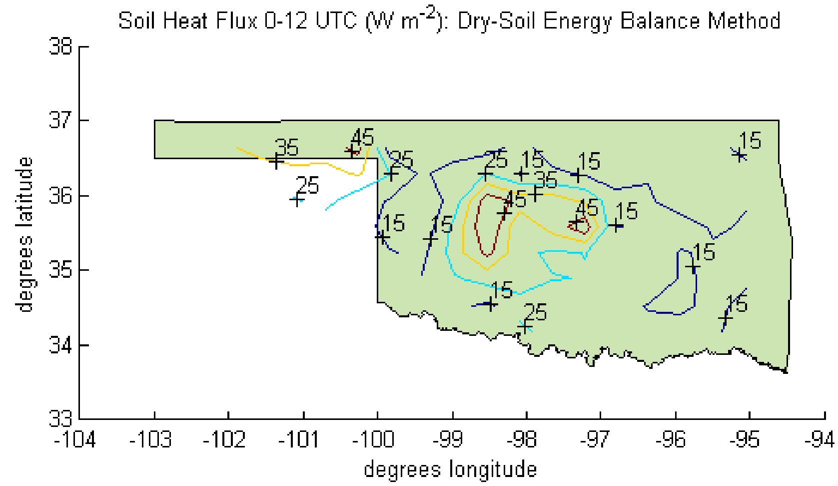

Figure 4: Estimate of time-mean surface heat flux, from 0-12 UTC on August 19, based on equation 3, using dry-soil heat capacity, and no changes in water content. Soil temperature data from roughly 80 of the Mesonet sites are interpolated to a 20 by 10 grid and the energy-balance heat flux is calculated on the grid each hour from 0-12 UTC, then averaged.

between residual and saturated, we can use measured changes in FWI at 5 and 25-cm depths to infer $\Delta c_i \approx c_w(\mathrm{vsm}_{\mathrm{sat}} - \mathrm{vsm}_{\mathrm{res}})\Delta\mathrm{FWI}$ (this method is admittedly somewhat dicey, as the relation between FWI and vsm is nonlinear (Illston et al., 2008)). Since most of the rain falls through an extremely humid boundary layer, we also assume $T_{\mathrm{rain}} \approx T_{\mathrm{air}}$. Together, these assumptions give us values for the surface flux that are slightly higher than the dry-soil estimate (Figure 5). The consideration of variable water content apparently has little effect under these assumptions. The surface heat flux may be overestimated in this framework, because in most cases, the precipitation rate substantially exceeds changes in soil moisture as reflected by changes in FWI. This suggests that either surface runoff is occurring, water is infiltrating below 50 cm, or that the nonlinear relationship between FWI and soil moisture is significantly biasing our results.

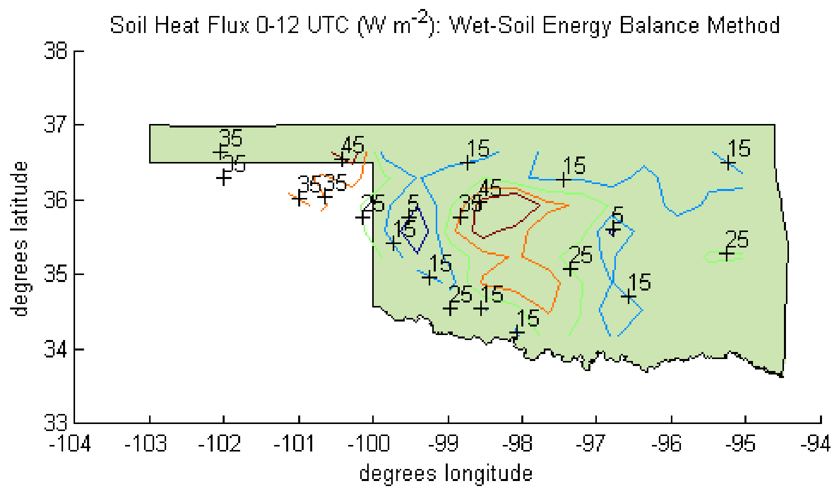

Figure 5: Estimate of time-mean surface heat flux, from 0-12 UTC on August 19, based on equation 4. Gridding methodology is as described in preceding figures.

For the above reasons, and also because attempts to model changes in soil moisture based on precipitation often give increases, rather than decreases, in soil heat content (implying negative surface fluxes) over part of the time period of consideration, we believe that the surface fluxes are probably not higher in a time-mean sense than indicated in either figures 4 or 5. However, a storm-centered composition of surface fluxes as a function of distance from the storm reveals somewhat different picture. Most of the high-flux meansurements are found close to the storm center, and few further away (Figure 6). Binning the data in 10-km radial intervals reveals that the mean surface flux 20-30 km from the storm center, as inferred by the dry-soil energy-balance method, is $\sim$160 W m^{-2}, though the standard deviation of the measurements is extremely large and estimates by the conductive-method flux are much lower (Figure 7). Much of the discrepancy between the two methods can be reconciled by adding the dry-soil energy-balance loss from the uppermost layer (7.5 cm) to the conductive flux estimate, but differences remain in the 20-30 km radial band where dry-soil energy-balance fluxe estimates are the highest. The accuracy of our methodology thus remains a somewhat unanswered question. In any case, if taken at face value, the dry-soil energy-balance surface fluxes are roughly 80 W m^{-2} when averaged over the closest 40 km to the reintensifying vortex. Soil measurements thus imply surface heat fluxes no higher than tens of W m^{-2} in a eulerian mean sense, but values may be well in excess of 100 W m^{-2} near the center of the vortex in a storm-following reference frame.

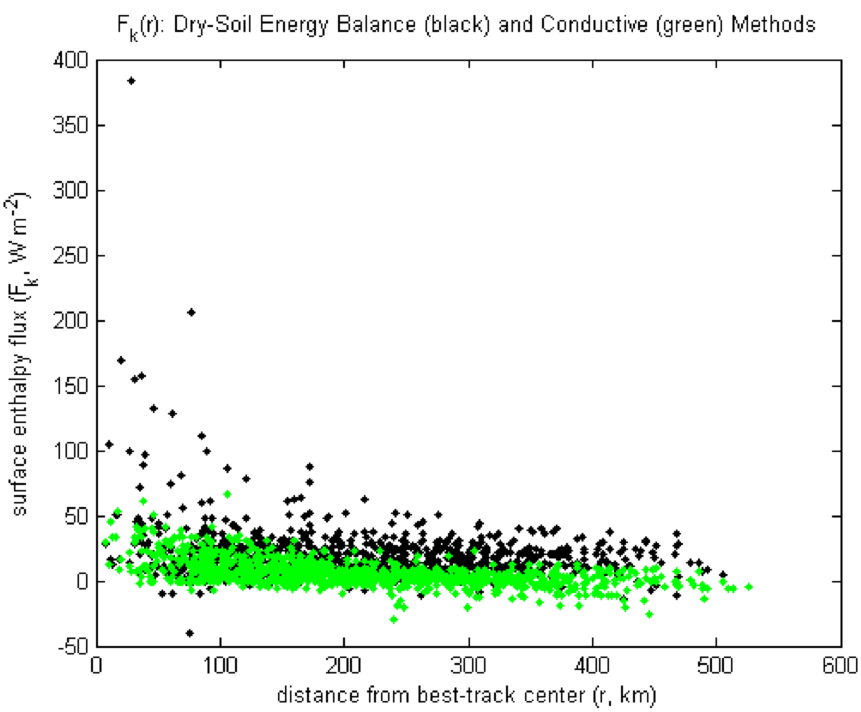

Figure 6: Estimate of the surface enthalpy flux as a function of distance from the best-track storm center (as defined in Knabb (2008)), from 0-12 UTC on August 19. Both the dry-soil energy balance (black) and conductive (green) methods are used, with fluxes calculated for each site on an hourly basis. A number of site-hours near the storm center have fluxes that exceed 100 W m^{-2} (963 total site-hours included).

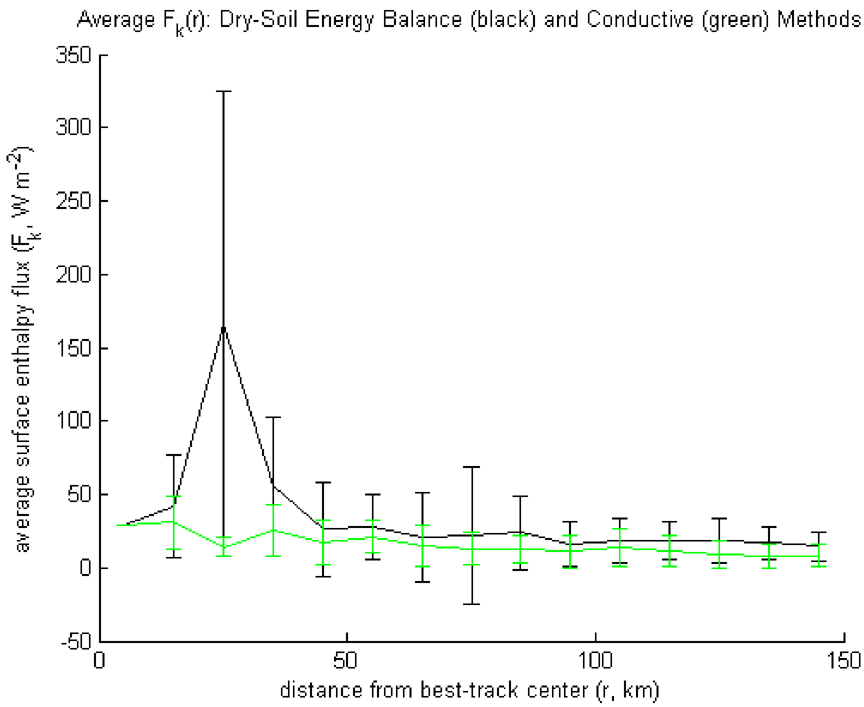

Figure 7: As in figure 6, but values in 10-km radial intervals are binned, and the mean value (solid line) and standard deviation (error bars) are shown instead of all site-hours. The conductive (green) and dry-soil energy-balance (black) methods agree fairly well, except for in the 20-30 km radial bin, where the dry-soil energy-balance method yields a result some 8 times larger than the conductive method.

3 Surface-Flux Reintensification

In the model of a TC as a steady dissipative heat engine, energy input by surface fluxes (multiplied by a thermodynamic efficiency factor) balances energy dissipation by surface drag (Bister and Emanuel, 1998). Equation 21 of Bister and Emanuel (1998) can be rearranged to give an equation for potential intensity (v_p) as based on surface enthalpy fluxes (F_k):

$$v_p^3 = \frac{F_k}{c_D}\frac{T_s - T_o}{T_o}, \tag{5}$$

where c_D is the drag coefficient, T_s the surface temperature, and T_o the outflow temperature. For a relatively smooth land surface with $c_D \sim 0.0015$ and a favorable thermodynamic environment with an efficiency of around 0.5 (e.g. surface temperature of 300 K and outflow temperature of 200 K), a minimal tropical storm with $v_p = 17.5$ m s^{-1} requires an extremely minimal surface enthalpy flux – only $F_k \sim 16$ W m^{-2}. It may seem surprising that such weak surface fluxes are consistent with a fairly strong surface circulation, and we have performed several simulations to explore reintensification of a warm-core cyclone via weak surface fluxes.

We use a modified version of the CHIPS model (Emanuel, 1995); the code is modified so that surface fluxes can be set to a constant value everywhere (or inside a certain radius, or outside a certain radius), instead of depending on wind speed and the surface thermodynamic disequilibrium. The first (perhaps surprising) finding is that spatially constant surface fluxes underneath a vortex that is weak but saturated in the free troposphere can result in rapid reintensification (Figure 8). Spatially constant surface fluxes translate into a greater amount of convective heating near the core of the cyclone, due to the higher precipitation efficiency there, rapidly amplifying the vortex. We say that such a case represents "reintensification," rather than simply "intensification," because the vortex is initialized with a saturated warm-core; the initial stages of formation of a TC likely lead to a saturated cold-core cyclone which only later can become a warm-core storm by intense convective heating.

The cases in Figure 8 have central pressure drops of ∼4 hPa /12 hours for the blue curves ($F_k = 40$ W m^{-2}), and ∼9 hPa /12 hours for the red curves ($F_k = 80$ W m^{-2}). These simulations have windspeeds that are only about two thirds of their potential intensity for the model thermodynamic environment, drag coefficients, and aforementioned surface fluxes, but nonetheless illustrate that rapid reintensification is quite possible for an axisymmetric vortex with an initially saturated core, even if surface fluxes are weak (by the standards of below-eyewall values in a mature oceanic TC) and spatially homogeneous.

As noted in the figure caption, the rate of reintensification of the vortex depends strongly on some parameters. One of these is the core saturation radius – values that are much larger or smaller than the value of 60 km chosen here give less dramatic results. Also, the fact that the core is saturated, and not just very humid, matters quite a bit (at least given the model formulation of precipitation efficiency) – if the core relative humidity is 95% instead of

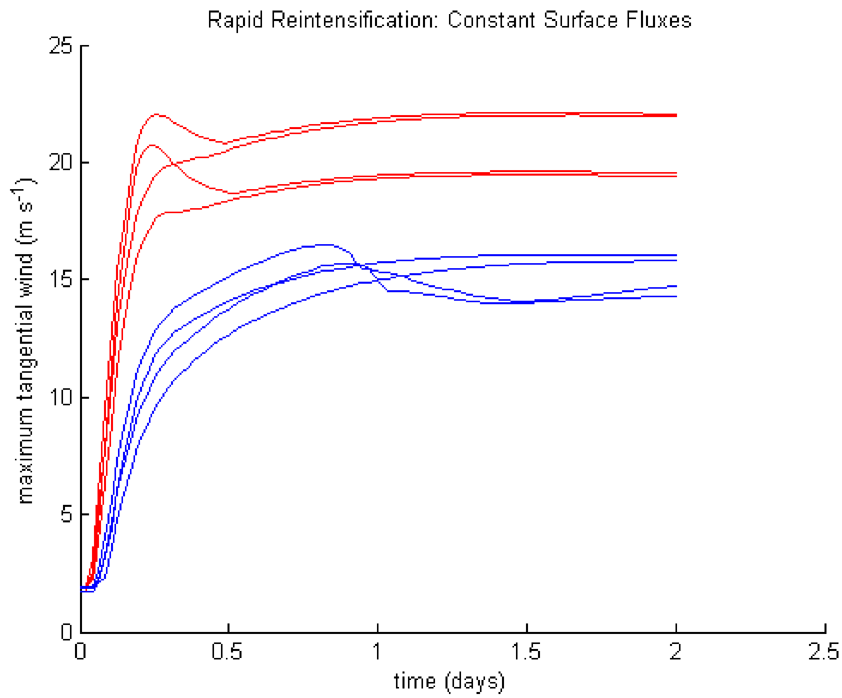

Figure 8: Reintensification of a weak warm-core vortex due to spatially constant surface fluxes of 40 W m^{-2} (blue) and 80 W m^{-2} (red). The initial vortex has maximum surface winds of 2 m s^{-1}, $r_0 = 350$ km, and a core radius of free-troposphere saturation of 60 km (reintensification is sensitive to this parameter, which has been chosen to give rapidly strengthening storms). Reintensification is relatively insensitive to the initial radius of maximum winds – 4 curves in each color have initial radius of maximum winds at 20, 40, 80, and 160 km.

100%, reintensification is dramatically slower, and the rapid-strengthening phase from about 0-0.25 days in Figure 8 is replaced by a much slower evolution. We present constant surface fluxes everywhere as a limiting case for simplicity, but rapid reintensification also occurs if surface fluxes are constant inside a certain radius, and zero outside (the reverse case, with zero surface fluxes inside a given radius and constant fluxes outside, does not rapidly reintensify). While a substantial number of assumptions go into these results, we believe they are illustrative of a general result, which is that even weak surface fluxes underneath a weak but saturated warm-core cyclone can result in a rapid pressure drop and increase in surface windspeed.

In relation to the reintensification of Erin, our simulations with a modified version of the CHIPS model suggest that surface enthalpy fluxes of 80 W m^{-2} might be able to roughly account for the observed dynamics – the rise in surface windspeed and drop in pressure for the red curves in Figure 8 are comparable to observations. We have shown on an Eulerian mean basis that surface fluxes are much weaker than this, but in a storm-centered reference frame surface fluxes could easily be this large when averaged over the 40-50 km nearest the storm center. Simulations by Evans et al. (2010) and Evans et al. (2011) indicate surface latent heat fluxes of up to 150 W m^{-2} within a somewhat larger domain, perhaps 50-100 km in radius (though they do not report the magnitude or direction of the sensible heat flux). Despite the sensitivity of CHIPS to model parameters, and the fact that the surface enthalpy fluxes are fairly weak relative to those in a mature oceanic TC, we believe that the results above provide significant support for surface-flux driven reintensification of TC Erin.

4 Discussion

Based on soil temperature measurements from the Oklahoma Mesonet, we have shown that surface enthalpy fluxes may have been as large as $\sim$ 100 W m^{-2} near the reintensifying core of TC Erin. Furthermore, intensity theory shows that such surface fluxes are consistent with surface winds comparable to observations, and an idealized axisymmetric model shows that rapid deepening of a weak warm-core vortex can be achieved with surface fluxes of this strength. There are, however, several major challenges to the theory that surface fluxes were responsible for reintensifying TC Erin.

First, we have discussed the role of surface fluxes in an idealized axisymmetric model, but the storm itself was highly asymmetric over the reintensification period, with a spiral band of deep convection stretching to the south and east, but little organized deep convection on the north and west (Figure 1). The dynamics of symmetrization thus may be a major issue for Erin – can a vortex with such pronounced convective asymmetry reintensify by the proposed pathway? We have also not assessed the degree to which surface fluxes are axially symmetric in a vortex-centered reference frame, though the large amount of scatter in Figures 6 and 7 suggests that might be somewhat difficult to ascertain with confidence from the existing data.

Also more troublingly asymmetric are values of surface equivalent potential temperature (θ_e). The storm reintensifies in an environment with very strong time mean gradients of surface θ_e (Figure 9). Differences on the order of 10 K are obvious between the southeast and northwest of the storm, though the cold (and/or dry) pool to the northwest of the storm, which appears partway through the time period of investigation, presumably forms due to the convective downdrafts from the storm itself (it is not pre-existing). While convection is most active in the southeast quadrant of the storm, it is not necessarily collocated with the highest values of surface θ_e, indicating conditional instability, which is supported by a sounding from the area (Figure 10). Furthermore, surface parcels in Figure 10 are not the ones with the highest equivalent potential temperature – such values are found a few hundred meters above the surface – posing a potential challenge to whether the storm's convective activity is even surface-based.

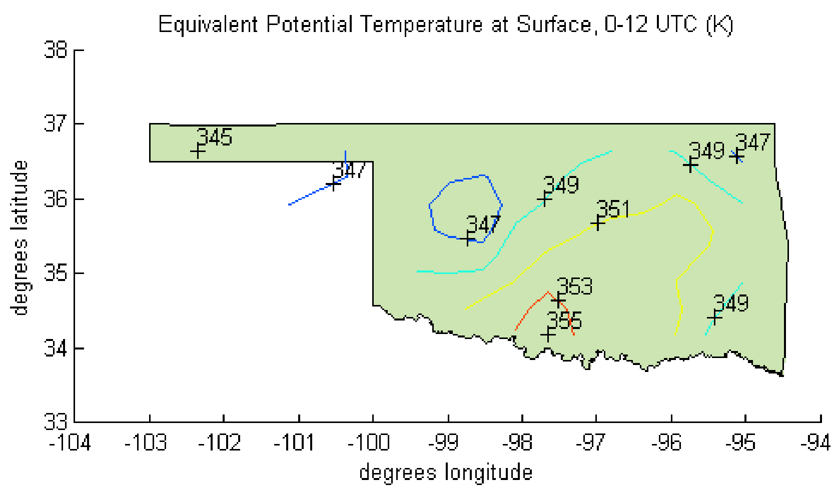

Figure 9: Plot of surface values of equivalent potential temperature, from 0-12 UTC on August 19. The highest values of θ_e are found substantially to the southeast of where the system reintensifies.

In order to compare the role of surface fluxes and conditional instability more directly, we can use the t_{CAPE} metric of Linders and Saetra (2010). They suggest that the importance of conditional instability relative to surface fluxes should be assessed by estimating the timescale it would take for surface fluxes to heat or moisten the boundary layer enough to produced observed CAPE values in the absence of convection. Using a CAPE value of 1800 $\mathrm{m}^{-2}\ \mathrm{s}^{-2}$, a surface layer mass of 500 kg, surface fluxes of 100 W m^{-2}, and a convective height of 10000 m in Equation 7 of Linders and Saetra (2010) gives $t_{\mathrm{CAPE}} \approx 27000$ s, or 7.5 hours. As this is comparable to the timescale of the reintensification, it suggests that ambient conditional instability may be a significant supply of energy to the system compared to surface fluxes. Though consideration of real (entraining) cumulus convection would reduce the real path-integrated buoyancy below sounding CAPE values, it seems rash to write off conditional instability completely, especially due to the briefness of the reintensification.

Due to sharp surface θ_e variations, another source of potential energy to the storm is advection of moist static energy in the boundary layer. We can compare advection of moist static energy and surface fluxes on an areal basis by choosing a surface layer thickness over which the surface winds and equivalent potential temperature apply – we will again use a boundary layer mass scale of 500 kg (Figure 11). We see that the advective flux of moist static energy is quite large compared to surface fluxes, at least in an eulerian-mean sense, with values over 150 W m^{-2} in a substantial part of the domain (we have not yet constructed an estimate of the storm-centered advective fluxes). This indicates that advective heating and moistening of the subcloud layer near the core of the storm may be more important than surface fluxes from an energetic standpoint, though it is difficult to know with certainty how important advection is, due to lack of knowledge of the vertical structure (of both θ_e and winds) in the subcloud layer.

Though we have shown that surface fluxes could be responsible for reintensifying the remnants of TC Erin, a final troubling question remains if that is indeed what occurred: why doesn't this sort of event occur more frequently? There are a number of plausible answers to this question – one of which is that we simply don't look for it hard enough. Even in the (seemingly quite clear-cut) case of Erin, the NHC did not ultimately choose to reclassify the reintensification as "tropical." More ambiguous situations that, for instance, lack clear eye formation, or that do not occur over such a meteorologically well-sampled domain as the Oklahoma Mesonet might be simply

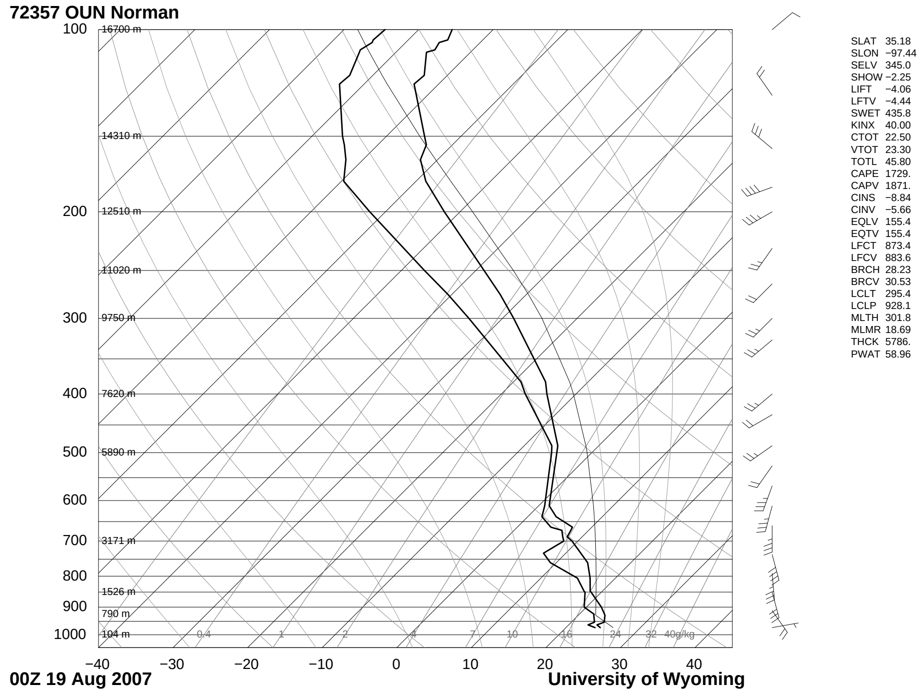

Figure 10: Sounding from Norman, Oklahoma at 00 UTC on August 19th. The sounding is conditionally unstable, with $\sim$ 1800 m^{-2} s^{-2} of CAPE, and very weak convective inhibition. From University of Wyoming sounding archive: http://weather.uwyo.edu/upperair/sounding.html

overlooked.

Another possible answer is that such events are rare simply due to the extreme sensitivity of reintensification to combined aspects of the residual vortex and land surface. This is especially apparent in CHIPS in regard to free tropospheric relative humidity of a weak vortex – plausible surface fluxes under a vortex over land essentially require that the vortex have a saturated core in order to rapidly reintensify. Also, there is the seemingly obvious point that in order for a vortex to reintensify to a tropical storm, it must first weaken considerably. Circumstances that both allow a vortex to weaken enough for reintensification to be discernible, and have low enough shear that the residual vortex can maintain its mid-level moisture, may be fairly rare. Rarer still would be such cases that also spend a significant amount

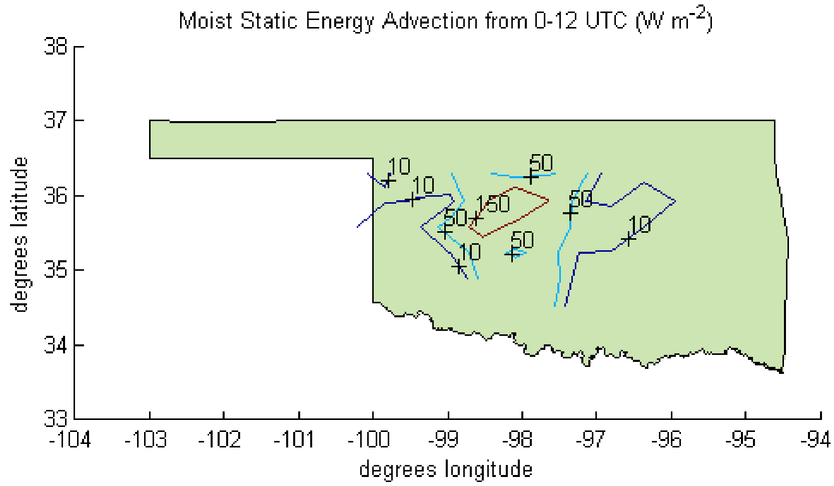

Figure 11: Estimate of time-mean advection of moist static energy ($Mc_p u \partial\theta_e/\partial x + Mc_p v \partial\theta_e/\partial y$), from 0-12 UTC on August 19 (W m^{-2}). Gridding methodology is as described in Figure 3.

of time over relatively smooth land with highly conductive soils, since most landfalling TCs spend more time over forests than plains, and many pass over much rougher terrain. Additionally, surface fluxes from warm soils may only be strong enough for reintensification in early-season storms (as the peak in soil temperature over land surfaces generally occurs before the peak in TC activity).

A third possibility, as suggested by Evans et al. (2010), is that conditional instability and/or interactions with an advective source of moist static energy (e.g. the low-level jet in this case) are necessary in order for Erin-like reintensification. This ought to be a testable hypothesis to some extent – other cases where asymmetric advection of moist static energy is significant, or where the environment is conditionally unstable, may exist for oceanic storms as well as landfalling ones.

5 Conclusions

In both basic theory, and in an axisymmetric model (under nearly ideal conditions), surface enthalpy fluxes of $\sim$ 80 W m^{-2} can result in rapid deepening of a weak warm-core vortex (e.g. the residue of a TC) at a rate and to a strength consistent with observations of the reintensification of Erin. Surface fluxes underneath Erin may have been as large as 100 W m^{-2}, but the amount of conditional instability present in the ambient environment suggests that consumption of CAPE could have been a comparable energy source to the storm on a timescale of hours – which is as long as the reintensification lasted. Furthermore, the advective flux of moist static energy from the southeast quadrant of the storm was likely substantially larger than surface enthalpy fluxes during the process of reintensification. Thus, although reintensification could have been theoretically driven by surface fluxes alone, the magnitudes of conditional instability and advection of moist static energy in this case, together with the extreme rarity of overland reintensification of tropical systems, suggest that surface fluxes were unlikely to be the sole player. But they may have been the key player – by triggering convection surface fluxes may have allowed for the additional release of CAPE; by initiating pressure falls, surface fluxes may have allowed for the convergence of high-entropy air into the core of the reintensifying cyclone. It is extremely difficult to convincingly demonstrate the role of any single physical process in a real meteorological event, and this case is no exception – multiple factors were almost certainly at play. Hopefully this paper has demonstrated that the answer to the question of Evans et al. (2010): "Whether latent heat fluxes on the order of 50-100 W m^{-2} can result in the TC-like intensification shown here" is an emphatic but qualified "yes," and we have laid the foundation for future work more thoroughly analyzing the energetics of the reintensification of Erin.

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