Tropical Transition of an Unnamed, High-Latitude, Tropical Cyclone over the Eastern North Pacific

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ABSTRACT

In early November 2006, an unnamed tropical cyclone (TC) formed via the tropical transition (TT) process at 42°N over the eastern North Pacific. An extratropical cyclone (EC), developing downstream of a thinning upper-tropospheric trough over the eastern North Pacific, served as the precursor disturbance that would ultimately undergo TT. The TT of the unnamed TC was extremely unusual—occurring over ~16°C sea surface temperatures in a portion of the eastern North Pacific basin historically devoid of TC activity.

This paper 1) identifies the upper- and lower-tropospheric features linked to the formation of the EC that transitions into the unnamed TC, 2) provides a synoptic overview of the features and processes associated with the unnamed TC’s TT, and 3) discusses the landfall of the weakening cyclone along the west coast of North America. As observed in previous studies of TT, the precursor EC progresses through the life cycle of a marine extratropical frontal cyclone, developing a bent-back warm front on its northern and western sides and undergoing a warm seclusion process. Backward air parcel trajectories suggest that air parcels isolated in the center of the transitioning cyclone were warmed in the lower troposphere via sensible heating from the underlying sea surface. Vertical cross sections taken through the center of the cyclone during its life cycle reveal its transformation from an asymmetric, cold-core, EC into an axisymmetric, warm-core, TC during TT. Ensemble reforecasts initialized after TT highlight the relatively low forecast skill associated with the landfall of the weakening cyclone.

1. Introduction

Tropical cyclones (TCs) are not exclusive to the tropics. While the environmental conditions deemed favorable for tropical cyclogenesis in the seminal works of Palmén (1948), Gray (1968), and DeMaria et al. (2001) are typically observed at tropical latitudes, environmental conditions can become favorable for tropical cyclogenesis in locations removed from the tropics. The global climatology of tropical cyclogenesis constructed by McTaggart-Cowan et al. (2013) reveals that the majority of TCs forming poleward of 25°N (25°S) in the Northern (Southern) Hemisphere during 1948–2010 developed in the presence of an upper-tropospheric disturbance (i.e., an upper-tropospheric low or trough) in a baroclinic environment (their Fig. 7). The frequency of formation of such TCs is shown to vary between and across individual ocean basins, likely in association with the frequency of upper-tropospheric disturbances intruding into these regions from the midlatitudes (e.g., Wernli and Sprenger 2007).

The TCs developing in the presence of an upper-tropospheric disturbance in a baroclinic environment typically form via the tropical transition (TT) process (Davis and Bosart 2003, 2004), during which an asymmetric, cold-core, extratropical cyclone (EC) transitions into an axisymmetric, warm-core TC. During the TT process, the EC often exhibits characteristic features of an evolving marine EC (i.e., bent-back warm front; Shapiro and Keyser 1990; Neiman and Shapiro 1993; Schultz et al. 1998; Hulme and Martin 2009a,b; Cordeira and Bosart 2011). In the initial stages of TT, vertical
wind shear in a baroclinic environment produces a region of upward motion that focuses deep convection and diabatic heating (Sutcliffe 1947). Vertical wind shear values are subsequently reduced by the redistribution of potential vorticity (PV) in the vertical due to differential diabatic heating (e.g., Hoskins et al. 1985; Davis and Emanuel 1991; Raymond 1992; Stoelinga 1996; Campa and Wernli 2012) and by divergent outflow in the upper troposphere, allowing the surface cyclone to intensify via air–sea interaction processes [i.e., wind-induced surface heat exchange (Emanuel 1986)].

Previous studies have established the importance of air–sea interactions, specifically sensible heat fluxes, in the formation of warm-core oceanic cyclones (e.g., Bosart and Bartlo 1991; Neiman and Shapiro 1993; Cordeira and Bosart 2011). A Lagrangian trajectory analysis performed by Cordeira and Bosart (2011) reveals that the warm seclusion and subsequent TT of the “Perfect Storm” (October 1991) in the northwestern North Atlantic involved the isolation of air parcels near the cyclone center that were warmed in the lower troposphere via sensible heating from the underlying Gulf Stream (their Fig. 9). The results of Cordeira and Bosart (2011) also reveal that the TT of the Perfect Storm occurred over sea surface temperatures (SSTs) below the traditional SST threshold for tropical cyclogenesis [26.5°C; e.g., Palmén (1948); Gray (1968)]. Recent studies by Mauk and Hobgood (2012) and McTaggart-Cowan et al. (2015) highlight the potential for tropical cyclogenesis to occur over SSTs < 26.5°C in environments characterized by reduced bulk column stability. This reduction of bulk column stability over SSTs < 26.5°C is typically associated with the presence of an upper-tropospheric disturbance that lowers the height of the dynamic tropopause (DT), steepening environmental lapse rates and facilitating the development of deep convection that serves as a catalyst for TT.

The TCs forming via TT have been documented in many basins where tropical cyclogenesis events occur annually, including the western North Atlantic (e.g., Moore and Davis 1951; Bosart and Bartlo 1991; Bracken and Bosart 2000; Davis and Bosart 2001; McTaggart-Cowan et al. 2006a; Evans and Guishard 2009; Guishard et al. 2009; Hulme and Martin 2009a,b), the western North Pacific (e.g., Wang et al. 2008), and the western South Pacific (e.g., Garde et al. 2010; Pezza et al. 2014). The TCs forming via TT have also been documented in basins where tropical cyclogenesis events are extremely rare, including the eastern North Atlantic (e.g., Case 1990; Franklin 2006; Beven 2006), the western South Atlantic (e.g., Pezza and Simmonds 2005; McTaggart-Cowan et al. 2006b; Evans and Braun 2012; Gozzo et al. 2014), and the Mediterranean Sea (e.g., Ernst and Matson 1983; Pytharoulis et al. 1999; Reale and Atlas 2001; Emanuel 2005; McTaggart-Cowan et al. 2010). In early November 2006, an unnamed TC [designated “Invest 91C” by the Central Pacific Hurricane Center (CPHC)] developed at 42°N in the eastern North Pacific basin (Fig. 1a). An EC, that formed downstream of a
thinning upper-tropospheric trough over the eastern North Pacific, served as the precursor disturbance that would ultimately undergo TT. The TT of Invest 91C, which took place between 0000 UTC 28 October and 0000 UTC 2 November 2006, occurred over ~16°C SSTs (Fig. 1b) far removed from tropical latitudes. The remnants of Invest 91C would ultimately make landfall along the extreme northwest (southwest) coast of Washington (British Columbia) at approximately 1600 UTC 3 November 2006, with measured wind gusts at Destruction Island, Washington, in excess of 29 m s⁻¹. This paper documents the formation of the EC that eventually became Invest 91C, the TT of this EC, and the subsequent evolution of the TC over the eastern North Pacific Ocean.

The remainder of this paper is organized as follows. The data and methodology used to analyze the development, TT, and landfall of Invest 91C are described in section 2. The upper- and lower-tropospheric features linked to the formation of the EC that transitioned into Invest 91C are identified and discussed in section 3. A synoptic overview of the life cycle of Invest 91C is presented in section 4 in order to document the features and processes associated with its TT and landfall. The diabatic processes and air–sea interactions occurring during the TT of Invest 91C are discussed in section 5. Key findings, conclusions, and a discussion of the rarity of this event are contained in section 6.

2. Data and methodology

The National Centers of Environmental Prediction Climate Forecast System Reanalysis (NCEP CFSR;
Saha et al. (2010) gridded dataset, with 0.5° × 0.5° horizontal grid spacing and 6-h temporal resolution, constitutes the primary data source for analyzing the development, TT, and landfall of Invest 91C. The 0.5° NCEP CFSR dataset is the first reanalysis dataset to be created using a global coupled atmosphere–ocean–land surface–sea ice model and to assimilate satellite radiances over the entire period of its availability.
making it well suited for analyzing the TT of an oceanic cyclone in a region with relatively sparse data coverage.

The upper- and lower-tropospheric features linked to the formation of the EC that transitioned into Invest 91C are examined every 24 h from 0000 UTC 24 October to 0000 UTC 28 October 2006 (hereafter all dates are in 2006) in section 3. The upper-tropospheric features deemed important to the formation of the EC that transitioned into Invest 91C are summarized using a Hovmöller diagram (Hovmöller 1949) of 250-hPa meridional winds plotted every 6 h from 0000 UTC 22 October through 0000 UTC 30 October. The standardized anomalies presented in the Hovmöller diagram, used to assess the characteristics of the 250-hPa meridional wind field, are computed from a long-term (i.e., 1979–2009) climatology derived from the 0.5° NCEP CFSR dataset using the methodology of Wilks (2011, his section 3.4.2). The structure of the large-scale flow pattern over the North Pacific and North America prior to and during the TT of Invest 91C is examined using daily standardized Pacific–North American (PNA) teleconnection indices (e.g., Wallace and Gutzler 1981; Barnston and Livezy 1987; Feldstein 2002). Daily standardized PNA teleconnection indices are calculated from 2.5° × 2.5° gridded NCEP–National Center for Atmospheric Research (NCAR) reanalysis (Kalnay et al. 1996; Kistler et al. 2001) 500-hPa geopotential height data between 24 October and 3 November following the methodology of Archambault et al. (2008, see their section 2a).

Figure 2a indicates the position of the center of Invest 91C every 6 h from 0000 UTC 28 October (genesis as a weak EC) to 1800 UTC 3 November (shortly after landfall). The position of the center of Invest 91C was identified using the minimum mean sea level pressure (MSLP) value associated with the cyclone within the 0.5° NCEP CFSR dataset (Fig. 2b). The colored dots in Fig. 2a correspond to the time periods analyzed in the synoptic overview of the life cycle of Invest 91C presented in section 4. Vertical cross sections, taken through the center of Invest 91C at 0000 UTC 29 October (red dot in Fig. 2a) and 0000 UTC 2 November (dark blue dot in Fig. 2a), are also presented in section 4 to illustrate the cyclone’s transition from an asymmetric, cold-core EC into an axisymmetric, warm-core TC.

As mentioned in section 1, deep convection and diabatic heating are required for the vertical redistribution of PV to occur during TT (Davis and Bosart 2003, 2004). Previous observational studies of TCs have identified regions of deep convection using satellite-derived infrared (IR) brightness temperature data (e.g., Romps and Kuang 2009; Hulme and Martin 2009a,b; Cordeira and Bosart 2010, 2011; Monette et al. 2012). Similarly, the present study utilizes IR brightness temperature data, obtained from the NCEP Climate Prediction Center (CPC) 4-km global (60°N–60°S) IR dataset (Janowiak et al. 2001; CPC 2007), in section 4 to identify regions of deep convection during the life cycle of Invest 91C. In addition, the present study utilizes lower-tropospheric thermal vorticity to represent the lower-tropospheric thermal structure of Invest 91C during its life cycle. Lower-tropospheric thermal vorticity (\(\zeta_T\)) will be defined in the present study as

\[
\zeta_T = \zeta_{g500} - \zeta_{g925},
\]

where \(\zeta_{g500}\) is 500-hPa geostrophic relative vorticity and \(\zeta_{g925}\) is 925-hPa geostrophic relative vorticity. A negative (positive) value of \(\zeta_T\) collocated with the center of Invest 91C indicates that the cyclone is warm (cold) core in the lower troposphere.

Near-surface weather data associated with the landfall of Invest 91C along the extreme northwest (southwest) coast of Washington (British Columbia) are presented in section 4, with data obtained from the National Data Buoy Center (NDBC) online data archive (NDBC 2006). The SST values displayed in Fig. 1b and discussed in sections 5 and 6 were obtained from analyses of daily 0.25° optimum interpolation SST
(OISST) data (Reynolds et al. 2007), collected using the Advanced Very High Resolution Radiometer (AVHRR-only). Finally, version 2 of the Earth System Research Laboratory/Physical Sciences Division Global Ensemble Forecast System (GEFS) reforecast dataset (Hamill et al. 2013) are used to evaluate the skill associated with the 42-h forecast of the landfall of Invest 91C. The GEFS reforecast dataset, created using the 2012 version of the NCEP GEFS, includes forecasts from an 11-member ensemble initialized once daily (0000 UTC) during the period from 1985 to the present. The MSLP forecast field, used to identify the track and intensity of Invest 91C, is available with \( \sim 0.5^\circ \times \sim 0.5^\circ \) horizontal grid spacing and is displayed every 6 h for consistency with the 0.5° NCEP CFSR dataset.

3. Upper- and lower-tropospheric precursors

The upper- and lower-tropospheric features linked to the formation of the EC that ultimately transitions into Invest 91C begin to interact over the extratropical North Pacific on 24 October. A meridionally confined 1000–500-hPa thickness gradient, located between a 996-hPa surface cyclone; \( \sim 350 \) km south of the Aleutian Islands and a 1036-hPa surface anticyclone; \( \sim 1400 \) km west of California, spans the North Pacific at 0000 UTC.
24 October (Fig. 3a). This zonally elongated baroclinic zone is associated with a $>70 \text{ m s}^{-1}$ upper-tropospheric jet that extends from northern Japan to western Canada (Figs. 3a,b). A broad surface low (EC1) begins to develop in the equatorward entrance region of the upper-tropospheric jet at 0000 UTC 24 October, downstream of an upper-tropospheric disturbance located over the Sea of Japan (Figs. 3a,b). Deep convection over and to the west of EC1, implied by the presence of upper-tropospheric divergent outflow emanating from a region of 600–400-hPa layer-averaged ascent (Fig. 3b), helps to reduce MSLP values associated with the cyclone over the following 24 h (Figs. 3a,c).

A second EC over southeastern Russia (EC2), located downstream of an upper-tropospheric trough in the 300–200-hPa layer-averaged PV field, begins to perturb the western portion of the North Pacific waveguide by 0000 UTC 25 October (Figs. 3c,d). Warm air advection (WAA) to the east of EC2 is associated with the amplification of the downstream ridge in the upper troposphere (Fig. 3d). Upper-tropospheric divergent outflow, emanating from a region of 600–400-hPa layer-averaged ascent collocated with the center of EC2, results in negative PV advection in the upper troposphere (e.g., Archambault et al. 2013, 2015) that slows the eastward progression of the upstream trough, contributes to rapid downstream ridge amplification, and enhances northwesterly flow downstream of the ridge axis between 0000 UTC 24 October and 0000 UTC 25 October (Figs. 3b,d).

Both EC1 and EC2 deepen and move northeastward between 0000 UTC 25 October and 0000 UTC 26 October (Figs. 3c,e). A corridor of WAA to the east of EC1 and EC2 extends from $\sim 40^\circ$ to $\sim 60^\circ$N by 0000 UTC 26 October (Fig. 3e), further amplifying the downstream
ridge. Negative PV advection by the 300–200-hPa layer-averaged irrotational wind continues to contribute to rapid ridge amplification and enhances meridional flow downstream of the ridge axis between 0000 UTC 25 October and 0000 UTC 26 October (Figs. 3d,f). Enhanced meridional flow downstream of the ridge axis is associated with the formation and amplification of a positively tilted upper-tropospheric trough over the central North Pacific by 0000 UTC 26 October (Fig. 3f), inferred from the orientation of PV contours in the 300–200-hPa layer-averaged PV field.

The development of the positively tilted upper-tropospheric trough over the central North Pacific coincides with the formation of a narrow corridor of relatively low MSLP values over the eastern North Pacific at 0000 UTC 26 October, in the equatorward entrance region of the North Pacific jet (Fig. 3e). This narrow corridor of relatively low MSLP values amalgamates into a ~1012-hPa closed low (L) over the following 24 h as the positively tilted upper-tropospheric trough amplifies upstream (Figs. 3e,g). Negative PV advection by the 300–200-hPa layer-averaged irrotational wind on the east side of the positively tilted upper-tropospheric trough causes the trough to slow, stretch, and thin over the eastern North Pacific between 0000 UTC 26 October and 0000 UTC 27 October (Figs. 3f,h). Continued stretching and thinning occurs over the following 24 h, coinciding with the equatorward movement of L to 40.5°N, 147.5°W by 0000 UTC 28 October (Figs. 3g–j).

The amplification of the North Pacific waveguide that results in the formation of L is summarized in a Hovmöller diagram of 250-hPa meridional winds averaged between 40° and 60°N (Fig. 4). Figure 4 highlights the eastward propagation of a Rossby wave train (e.g., Namias and Clapp 1944; Chang 1993; Orlanski and Sheldon 1993, 1995; Hakim 2003; Danielson et al. 2004) across the North Pacific basin between 0000 UTC 24 October and 0000 UTC 28 October. The upper-tropospheric trough (T1) upstream of EC2 is located over eastern Asia at 0000 UTC 24 October. Figure 4 reveals that T1 propagates across the Asiatic continent between 0000 UTC 22 October and 0000 UTC 24 October as part of a preexisting Rossby wave. The amplification of a preexisting Rossby wave has been shown to precede TT in numerous cases in the North Atlantic basin (e.g., Bosart and Bartlo 1991; Cordeira and Bosart 2010, 2011; Avila 2012), resulting in the downstream perturbation of the midlatitude waveguide and the formation of the

![Graph](image-url)
upper-tropospheric disturbance required for TT to occur (Davis and Bosart 2003, 2004).

Upper-tropospheric divergent outflow emanating from regions of deep convection associated with EC1 and EC2 (Fig. 3b), as well as WAA occurring to the east of each cyclone (Fig. 3a), helps to rapidly amplify the downstream ridge (R1) between 0000 UTC 24 October and 0000 UTC 25 October (Fig. 4). Enhanced northerly flow downstream of R1 over the following 24 h is associated with the formation and amplification of an upper-tropospheric trough over the central North Pacific (T2). The magnitude of the northerly and southerly 250-hPa meridional winds associated with R1 and T2 continue to increase between 0000 UTC 26 October and 0000 UTC 27 October (Fig. 4), coinciding with the tightening of the upper-tropospheric PV gradients in the central and eastern North Pacific (Figs. 3f,h). This gradient tightening is attributed to negative PV advection by the irrotational wind in the upper troposphere (Figs. 3f,h), which has been shown by Archambault et al. (2013, 2015) to result in PV frontogenesis and an enhancement of the upper-tropospheric jet.

Negative PV advection by the irrotational wind on the western and eastern sides of T2 (Figs. 3f,h,j) causes T2 to continue to amplify and thin between 0000 UTC 27 October and 0000 UTC 28 October (Fig. 4). Figure 4 reveals northerly (southerly) 250-hPa meridional winds in excess of $-2\sigma$ ($+2\sigma$) upstream (downstream) of the trough axis during this period, suggesting that the upper-tropospheric trough associated with the formation of L is an anomalous feature in the eastern North Pacific basin in late October. An examination of daily standardized PNA teleconnection indices in late October indicates that the anomalous amplification of
the midlatitude waveguide over the eastern North Pacific corresponds to a transition of the PNA from neutral to slightly negative (Fig. 2b), with values decreasing from $-0.102\sigma$ to $-0.998\sigma$ between 0000 UTC 24 October and 0000 UTC 27 October. Figure 2b reveals that daily standardized PNA teleconnection indices are not exceptionally anomalous prior to the formation of L on 0000 UTC 27 October, suggesting that the portion of the large-scale flow pattern described by the PNA is not required to be highly amplified for TT to occur in the eastern North Pacific.

4. Tropical transition of Invest 91C

a. Warm seclusion

An elongated region of low MSLP values (previously designated L), centered at 40.5°N, 147.5°W at 0000 UTC 28 October, exists on the warm side of a 925–500-hPa thickness gradient in a region of negative 925–500-hPa $\zeta_T$ values (Fig. 5a). The strip of 925–850-hPa layer-averaged relative vorticity associated with this region of negative $\zeta_T$ is located downstream of the thinning upper-tropospheric trough identified in section 3 (Fig. 5b). A meridionally elongated corridor of $>30$-mm precipitable water (PW) values extends poleward downstream of this thinning upper-tropospheric trough, collocated with a region of 600–400-hPa layer-averaged ascent (Fig. 5c). Upper-tropospheric divergent outflow to the west of the region of 600–400-hPa layer-averaged ascent tightens the upper-tropospheric PV gradient via negative PV advection and slows the eastward propagation of the southern portion of the thinning upper-tropospheric trough. The cloud field at 0000 UTC 28 October in Fig. 5d aligns nicely with the region of 600–400-hPa layer-averaged ascent in Fig. 5c.

Fig. 9. As in Fig. 5, but at 0000 UTC 31 Oct 2006.
Differential cyclonic vorticity advection occurring over the elongated region of low MSLP values (not shown) is associated with the formation of a 999-hPa EC over the following 24 h (Figs. 5a and 6a). A region of WAA [cold air advection (CAA)] begins to wrap cyclonically around the eastern (western) edge of the EC by 0000 UTC 29 October, resulting in a discernable warm (cold) front in the 925–500-hPa thickness and $z_T$ fields (Fig. 6a). The southern portion of the thinning upper-tropospheric trough has become cut off from the northern portion at this time (Fig. 6b), allowing the corridor of warm, moist, air and 600–400-hPa layer-averaged ascent in Figs. 5a and 5c to begin to wrap cyclonically around the northern edge of the EC (Figs. 6a,c). The diabatic processes contributing to the cutting off of the southern portion of the thinning upper-tropospheric trough by 0000 UTC 29 October will be discussed in greater detail in section 5.

A vertical cross section, taken along line A–A’ in Fig. 6d, highlights the different air masses wrapping cyclonically around the 999-hPa EC (Fig. 7). The narrow corridor of warm, moist, air to the east of the EC center in Figs. 6a and 6c is indicated by $>312$-K equivalent potential temperature ($\theta_e$) values and by relatively high potential temperature ($\theta$) values in the lower and midtroposphere immediately to the east of the MSLP minimum. A region of cold, dry air to the west of the EC center in Figs. 6a and 6c is indicated by $<306$-K $\theta_e$ values and by relatively low $\theta$ values in the lower and midtroposphere to the west of the MSLP minimum. Directly over the MSLP minimum, a $\sim18$-K decrease in $\theta_e$ from the surface to $\sim600$ hPa denotes a region of potential instability. A region of PV values $>2$ PVU...
(1 PVU = \(10^{-6}\) K kg\(^{-1}\) m\(^2\) s\(^{-1}\)) associated with the upper-tropospheric cutoff shown in Fig. 6b extends from the stratosphere to \(\sim\)600 hPa to the west of the EC center. A region of >2-PVU values also exists directly over the EC center in the lower troposphere, collocated with a lower-tropospheric frontal boundary.

The corridor of warm, moist air wraps around the northern and western edges of the EC by 0000 UTC 30 October, resulting in the bent-back warm front visible in the 925–500-hPa \(\zeta_T\) field (Fig. 8a). Negative \(\zeta_T\) values, which coincide with the 925–850-hPa layer-averaged relative vorticity maximum, have become vertically aligned with the upper-tropospheric cutoff (Figs. 8a,b). As will be discussed in section 5, the vertical alignment of the surface cyclone and upper-tropospheric cutoff between 0000 UTC 29 October and 0000 UTC 30 October reduces the bulk column stability near the EC center. Reduced bulk column stability and PW values >25 mm beneath the upper-tropospheric cutoff (Fig. 8c) creates a favorable environment for the development of convection near the EC center (Fig. 8d).

b. Tropical transition

Condensation heating from buoyancy-induced convection near the EC center, discussed more explicitly in section 5, allows the cyclone to begin to gain TC-like characteristics and lose its frontal structure by 0000 UTC 31 October. Negative \(\zeta_T\) values associated with the cyclone center have become removed from the rest of the bent-back warm front (Fig. 9a) and remain collocated with the upper-tropospheric cutoff at this time (Fig. 9b). A slight increase in DT \(\theta\) values over the cyclone center (Fig. 9b), corresponding to a 1-PVU
reduction in 300–200-hPa layer-averaged PV values over the cyclone center (Fig. 9c), suggests that the upper-tropospheric cutoff is being eroded by the diabatic redistribution of PV in the vertical. Midtropospheric ascent continues to occur to the north and east of the center of circulation in a region of >25-mm PW values (Fig. 9c), coinciding with the position of a spiral band observable in the IR satellite imagery (Fig. 9d).

The central pressure of the transitioning cyclone reaches 991 hPa by 0000 UTC 1 November (Figs. 2b and 10a). Values of $\zeta_T$ have decreased to $<-16 \times 10^{-3}$ s$^{-1}$ near the cyclone center at this time, with a reduction in $\zeta_T$ values observed throughout the majority of the analyzed domain (Figs. 9a and 10a). The 925–850-hPa layer-averaged relative vorticity maximum has moved northward in response to a progressive upper-tropospheric trough that wrapped cyclonically around the southern portion of the transitioning cyclone during the previous 24 h (Figs. 9b and 10b). Midtropospheric ascent is observed over and surrounding the center of the transitioning cyclone at this time (Fig. 10c). Furthermore, a small region of deep convection (indicated by cold cloud tops in the IR satellite imagery in Fig. 10d) is collocated with the cyclone center, corresponding to the analyzed regions of midtropospheric ascent shown in Fig. 10c.

The TC-like characteristics of the transitioning cyclone at 0000 UTC 1 November (Figs. 10a–d) caused CPHC to label the storm “Invest 91C” at 1200 UTC 1 November. The central pressure of Invest 91C decreases from 991 to 989 hPa between 0000 UTC 1 November and 0000 UTC 2 November (Figs. 10a and 11a), while 925–850-hPa layer-averaged relative vorticity values $>2.0 \times 10^{-4}$ s$^{-1}$ are maintained near the cyclone center (Figs. 10b and 11b). Midtropospheric ascent persists near the near cyclone center at 0000 UTC 2 November (Fig. 11c), collocated with >30-mm PW values and a region of deep convection (Figs. 11c,d).

A vertical cross section, taken along line B–B’ in Fig. 11d, indicates that Invest 91C has completely transitioned into an axisymmetric, warm-core TC by 0000 UTC 2 November (Fig. 12). The TC characteristics displayed in Fig. 12 are nearly identical to those displayed in Fig. 18 of Hulme and Martin (2009b), who first suggested the TT of Invest 91C in the eastern North Pacific. Relatively high $\theta$ values located over the center of Invest 91C, indicated by a bowing down of $\theta$ contours toward the surface throughout the depth of the troposphere, confirm the warm-core structure of the cyclone suggested in Fig. 11a. Vertically oriented $\theta_e$ contours surrounding the MSLP minimum suggest that deep convection has been occurring near the center of the cyclone. The upper-tropospheric PV maximum that extended from the stratosphere to $\sim$600 hPa at 0000 UTC 29 October (Fig. 7) no longer exists at 0000 UTC 2 November (Fig. 12), likely due to the diabatic redistribution of PV in the vertical associated with persistent deep convection. A
PV tower, also indicative of the diabatic redistribution of PV in the vertical, is located over the MSLP minimum between the surface and -400 hPa.

c. Landfall

Invest 91C begins to approach the west coast of North America between 0000 UTC 2 November and 0000 UTC 3 November, traveling 10° of longitude in 24 h (Fig. 2a). The minimum MSLP value associated with the center of the cyclone has increased from 989 to 993 hPa during this period as the MSLP field becomes broader and less organized (Figs. 11a and 13a). Values of $\zeta_T$, however, remain robust ($\sim -14 \times 10^{-5}$ s$^{-1}$) immediately to the south of the minimum in MSLP (Fig. 13a), corresponding to the position of the 925–850-hPa layer-averaged relative vorticity maximum (Fig. 13b). Figure 13c indicates that midtropospheric ascent is still occurring over Invest 91C at 0000 UTC 3 November, collocated with a region of >25-mm PW values (Fig. 13c). This midtropospheric ascent, however, is associated with minimal convective activity in the IR brightness temperature field (Fig. 13d). Deep-layer (850–200 hPa) vertical wind shear exceeding 18 m s$^{-1}$ over the cyclone center forces convection downshear of Invest 91C at this time and likely prevents further intensification.

The final advisory for Invest 91C was issued by CPHC at 1200 UTC 3 November, only a few hours prior to the landfall of the weakening disturbance. Figure 2a reveals that the remnants of Invest 91C traveled rapidly to the northeast between 0000 UTC 3 November and 1800 UTC 3 November, making landfall along the extreme northwest (southwest) coast of Washington (British Columbia) between 1600 and 1800 UTC 3 November. A region of >486-dam 925–500-hPa thickness values is located...
over extreme northwest (southwest) coast of Washington (British Columbia) at 1800 UTC 3 November (Fig. 14a), collocated with a region of 925–850-hPa layer-averaged relative vorticity values in excess of $1.5 \times 10^{-3}$ s$^{-1}$ (Fig. 14b). The strip of relatively high 925–850-hPa layer-averaged relative vorticity values along the Puget Sound is likely associated with surrounding orography, occurring between the Olympic Mountains and the Cascade Range at 0000 UTC 3 November and 1800 UTC 3 November (Figs. 13b and 14b). Midtropospheric ascent and convection associated with the remnants of Invest 91C are difficult to distinguish from surrounding features at 1800 UTC 3 November (Figs. 14c,d), making further synoptic-scale analysis difficult.

The sensible weather associated with the landfall of the remnants of Invest 91C can be analyzed using observations obtained from the NDBC online data archive (see section 2). Figure 15 reveals that pressure values at Destruction Island, Washington, decrease steadily from 1024 to 1000 hPa in the days leading up to the landfall of the remnants of Invest 91C (i.e., between 2000 UTC 31 October and 2200 UTC 2 November). Wind speed values remain relatively weak during the first portion of this period, but increase from 2.8 to 14.0 m s$^{-1}$ between 0200 UTC 2 November and 2200 UTC 2 November. Pressure (wind speed) values are shown to increase (decrease) gradually between 2200 UTC 2 November and 0800 UTC 3 November. However, at 0900 UTC 3 November, pressure (wind speed) values begin to fall (rise) rapidly at the Destruction Island lighthouse in association with the approaching remnants of Invest 91C. Wind speed and wind gust values of 26.2 and 29.2 m s$^{-1}$, respectively, are observed at 1600 UTC 3 November, corresponding to a pressure...
measurement of 998.4 hPa. Pressure (wind speed) values rise (fall) rapidly after the passage of the remnants of Invest 91C, indicating the relatively small areal extent of the disturbance.

The data recorded at the Destruction Island lighthouse is representative of wind speed, wind gust, and pressure measurements recorded at surrounding observation stations included in the NDBC online data archive. Table 1 indicates that the southern observation stations (Destruction Island and Cape Elizabeth, Washington) recorded minimum pressure values at 1600 UTC 3 November, 2 h earlier than the northern observation stations (Neah Bay and Tatoosh Island, Washington). The northern observation stations, however, recorded lower minimum pressure values than the southern observation stations, likely due to their alignment with the remnants of Invest 91C (Fig. 2a). Finally, the observation stations closest to the coast of Washington (Destruction Island and Tatoosh Island) recorded the fastest wind speeds and wind gusts of the observation stations included in Table 1, likely due to terrain channeling of previously unimpeded flow over the Pacific Ocean.

The landfall of Invest 91C, though not associated with excessive wind damage, coincided with the beginning of a 5-day period of heavy precipitation and extensive flooding in western Washington. The National Climatic Data Center Storm Events database (NCDC 2013; the 2006 data were released in 2013 according to https://www.ncdc.noaa.gov/stormevents/versions.jsp) reveals that 102–254 mm (254–965 mm) of precipitation fell in coastal lowlands (mountain ranges) during 3–7 November, resulting in two fatalities and ~$70 million in property damage. The landfall of Invest 91C, along with residual subtropical moisture in the eastern North Pacific associated with its TT (Fig. 14c), contributed to this prolonged period of heavy precipitation and the destruction that ensued.

5. Dynamical linkages and diabatic processes

As discussed in section 1, the potential exists for TT to occur over SSTs < 26.5°C in environments characterized by

TABLE 1. Measurements recorded by observation stations during the landfall of the remnants of Invest 91C on 3 Nov 2006. Measurements correspond to the time when the lowest pressure value was recorded at each observation station. Observation stations are listed from north to south.

<table>
<thead>
<tr>
<th>Station name (station No.)</th>
<th>Location</th>
<th>Time</th>
<th>Pressure (hPa)</th>
<th>Wind speed (m s⁻¹)</th>
<th>Wind gust (m s⁻¹)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Neah Bay, WA (46087)</td>
<td>48.5°N, 124.7°W</td>
<td>1800 UTC</td>
<td>995.5</td>
<td>15.9</td>
<td>21.1</td>
</tr>
<tr>
<td>Tatoosh Island, WA (TTIW1)</td>
<td>48.4°N, 124.7°W</td>
<td>1800 UTC</td>
<td>994.8</td>
<td>20.0</td>
<td>24.6</td>
</tr>
<tr>
<td>Destruction Island, WA (DESW1)</td>
<td>47.7°N, 124.5°W</td>
<td>1600 UTC</td>
<td>998.4</td>
<td>26.2</td>
<td>29.2</td>
</tr>
<tr>
<td>Cape Elizabeth, WA (46041)</td>
<td>47.3°N, 124.7°W</td>
<td>1600 UTC</td>
<td>999.1</td>
<td>14.4</td>
<td>18.1</td>
</tr>
</tbody>
</table>
by reduced bulk column stability. McTaggart-Cowan et al. (2015) utilized the coupling index (CI), defined by Bosart and Lackmann (1995) as the difference between $\theta$ on the DT and $\theta_e$ at 850 hPa, to approximate the bulk column stability associated with TCs forming in the presence of an upper-tropospheric disturbance during 1989–2013. The results of McTaggart-Cowan et al. (2015) suggest that a CI value of 22.5°C is the upper limit for TT-based TC developments globally. A quasi-Lagrangian time series of CI values and 925–500-hPa layer-averaged PV values, averaged within a 3° × 3° box centered over Invest 91C every 6 h between 0000 UTC 28 October and 1200 UTC 3 November, is depicted in Fig. 16. A 3° × 3° box was determined to be small enough to capture CI and 925–500-hPa layer-averaged PV values near the cyclone center, but large enough for the calculated values to be stable. Figure 16 reveals that CI values decrease from 23.8° to 2.6°C between 0000 UTC 28 October and 0000 UTC 30 October, as the upper-tropospheric disturbance and the surface cyclone become increasingly vertically aligned (Figs. 5–8). The CI value of 2.6°C at 0000 UTC 30 October is 19.9°C below the upper limit for TT-based TC developments identified in McTaggart-Cowan et al. (2015), suggesting that sufficient instability exists for TT to occur. In contrast to values of the CI, values of 925–500-hPa layer-averaged PV increase considerably between 0000 UTC 28 October and 0000 UTC 30 October. The maximum rate of increase occurs between 0600 UTC 28 October and 0000 UTC 29 October, suggesting condensation heating is occurring within regions of deep convection near the cyclone center and increasing (decreasing) PV values in the lower (upper) troposphere.

The condensation heating suggested in Fig. 16 is explored in greater detail using 48-h air parcel trajectories ending at 0000 UTC 29 October (Fig. 17). This Lagrangian diagnostic is used to assess the properties of air parcels experiencing slantwise ascent within the warm sector of Invest 91C, which may experience an increase (decrease) in lower-tropospheric (upper tropospheric) PV due to differential diabatic heating. Only air parcels ascending 600 hPa in 48 h and terminating within the 350–150-hPa layer to the north of Invest 91C were plotted in order to capture trajectories within the cyclone’s warm conveyor belt (WCB; e.g., Browning et al. 1973; Wernli and Davies 1997; Madonna et al. 2014). A large subset of air parcels in the WCB originate in the lower troposphere to the south of Invest 91C at 0000 UTC 27 October, move poleward over the following 30 h, and rapidly ascend from >900 to <300 hPa to the east of the cyclone between 0600 UTC 28 October and 0000 UTC 29 October (Fig. 17a). An additional subset of air parcels originate in the lower troposphere to the north of Invest 91C at 0000 UTC 27 October, move equatorward over the following 30 h, and rapidly ascending from >900 to <300 hPa to the east of the cyclone between 0600 UTC 28 October and 0000 UTC 29 October. Figures 17b–d indicate that air parcels experience a mean increase (decrease) in PV in the lower (upper) troposphere of ~0.75 PVU during their period of rapid ascent, coinciding with a mean increase (decrease) in $\theta$ (specific humidity) values of ~30 K (~9 g kg$^{-1}$). These patterns are indicative of condensation heating and precipitation occurring within the WCB (Madonna et al. 2014) and illustrate
that the enhancement of 925–500-hPa layer-averaged PV shown in Fig. 16 is associated with diabatic processes. The corresponding reduction in upper-tropospheric PV likely contributes to the cutting off the southern portion of the upper-tropospheric trough over the surface cyclone at this time (Figs. 6a,b).

Figure 16 reveals that CI values \( < 4.0^\circ C \) persist between 0000 UTC 30 October and 0000 UTC 2 November as the upper-tropospheric disturbance and surface cyclone remain vertically aligned (Figs. 8–11). Values of 925–500-hPa layer-averaged PV decrease slightly during this period, but remain relatively high compared to their minimum value at 0600 UTC 28 October (Fig. 16). This slight decrease is consistent with a gradual reduction in the strength and areal extent of deep convection surrounding the cyclone center between 0000 UTC 30 October and 0000 UTC 2 November (Figs. 8d–11d). Deep convection continues to weaken between 0000 UTC 2 November and 1200 UTC 3 November as Invest 91C rapidly moves toward land (Figs. 11d and 13d). In contrast to values of 925–500-hPa layer-averaged PV, which continue to decrease during this period, CI values begin to increase (Fig. 16). This increase in CI values is associated with an increase in the elevation of the DT over the cyclone center due to 1) differential diabatic heating occurring within regions of deep convection and the resulting reduction of upper-tropospheric PV (Fig. 17), as well as 2) the northeast movement of the upper-tropospheric disturbance away from the surface cyclone (Figs. 11–13).

In addition to condensation heating, sensible heat fluxes from the underlying sea surface have been shown to play an important role in the formation of warm-core oceanic cyclones (e.g., Bosart and Bartlo 1991; Neiman and Shapiro 1993; Cordeira and Bosart 2011). Figure 18 illustrates 48-h backward trajectories of 900–800-hPa air parcels located near the center of Invest 91C at 0000 UTC 30 October (during its warm seclusion). A large subset of air parcels originate in the lower troposphere to the north of Invest 91C at 0000 UTC 28 October, progress
equatorward, and gradually descend toward the sea surface over the following 30 h before ascending into the 900–800-hPa layer between 0600 UTC 29 October and 0000 UTC 30 October (Fig. 18a). These air parcel trajectories are nearly identical to the cold conveyor belt (CCB; e.g., Carlson 1980) trajectories associated with the warm seclusion of the Perfect Storm (Cordeira and Bosart 2011, their Fig. 9). An additional subset of air parcels originate at the sea surface near the center of Invest 91C after 0000 UTC 28 October, remaining below 900 hPa for ~18–36 h before ascending into the 900–800-hPa layer by 0000 UTC 30 October (Fig. 18a).

Figures 18b–d indicate that the mean PV, $\theta$, and specific humidity values of air parcels remaining near the sea surface increase by ~0.75 PVU, ~10 K, and ~4 g kg$^{-1}$, respectively, in the hours prior to Invest 91C’s warm seclusion. Eastern North Pacific SSTs near and to the north of the center of Invest 91C were ~1σ above the long-term (i.e., 1982–2011) climatological SST values calculated by Banzon et al. (2014) in the 48 h prior to 0000 UTC 30 October (not shown). The presence of anomalously high SSTs likely enhances sensible heat fluxes between the sea surface and lower-tropospheric air parcels during this period, steepening local lapse rates (e.g., Fig. 7), and allowing for persistent convection to occur near the cyclone center. Figures 18b–d reveal a slight increase (decrease) in PV and $\theta$ (specific humidity) values during the ~6 h of ascent prior to 0000 UTC 30 October, indicative of condensation heating and precipitation occurring near the cyclone center during this period (Madonna et al. 2014).

6. Discussion and conclusions

The TT of Invest 91C occurred at 42°N in the eastern North Pacific between 0000 UTC 28 October and 0000 UTC 2 November 2006. Despite the rarity of tropical cyclogenesis poleward of ~25°N in the eastern North Pacific (McTaggart-Cowan et al. 2013), the features and processes associated with Invest 91C’s transition from an asymmetric, cold-core EC into an axisymmetric, warm-core TC are similar to those found in previous TT studies. The key features and processes associated with the formation and TT of Invest 91C, as well as the rarity of the event, will be discussed in the remainder of this section.

a. Upper- and lower-tropospheric precursors

The upper-tropospheric trough associated with the formation and TT of Invest 91C develops over the
eastern North Pacific as part of an eastward-propagating Rossby wave train in late October 2006. Upper-
tropospheric divergent outflow and WAA occurring to
the vicinity of two ECs over eastern Asia facilitate the
amplification an upper-tropospheric ridge (part of a
preexisting Rossby wave) over the western North Pacific
between 0000 UTC 24 October and 0000 UTC 26 Oc-
tober. Enhanced northerly flow downstream of the
upper-tropospheric ridge results in the formation and
amplification of a positively tilted upper-tropospheric
trough, ultimately associated with the TT of Invest 91C,
over the central North Pacific by 0000 UTC 26 October.
Negative PV advection by the 300–200-hPa layer-
averaged irrotational wind on either side of the posi-
tively tilted upper-tropospheric trough, ultimately associated with the TT of Invest 91C,
over the central North Pacific by 0000 UTC 26 October.
A closed low, developing downstream of the
positively tilted upper-tropospheric trough during this
period (between $-0.102\sigma$ and $-0.998\sigma$). These results
suggest that the large-scale flow pattern described by the
PNA is not required to be highly amplified for TT to
occur in the eastern North Pacific, and that the location
and orientation of individual upper-tropospheric dist-
turbances may play a more important role in de-
termining TT potential. As in previous TT studies, the
amplification of a preexisting Rossby wave and the
subsequent perturbation of the downstream midlatitude
waveguide led to the formation of the upper-
tropospheric disturbance associated with the TT of In-
vest 91C. Additional research is required to determine
the percentage of TT events preceded by the amplifi-
cation of a preexisting Rossby wave and the various
predictability issues associated with different upstream
precursor configurations.

b. Warm seclusion and TT

Invest 91C was able to form over a region of $\sim$16°C
SSTs, considerably lower than the traditional threshold
for TC development, due to the presence of an upper-
tropospheric trough that lowered the height of the DT
and reduced deep-layer stability near the cyclone center
(e.g., Mauk and Hobgood 2012; McTaggart-Cowan et al.
2015). This study supports the findings of Davis and
Bosart (2003, 2004) and Hulme and Martin (2009a,b)
that an EC, developing in association with an upper-tropospheric trough approaching a lower-tropospheric baroclinic zone, serves as the precursor disturbance to TT. As suggested by Hulme and Martin (2009b) and Cordeira and Bosart (2011), the EC progresses through the life cycle of a marine extratropical frontal cyclone (Shapiro and Keyser 1990), developing a bent-back warm front on its northern and western sides prior to TT. A Lagrangian trajectory analysis of Invest 91C at 0000 UTC 30 October reveals that air parcels isolated near the cyclone center during its warm seclusion were likely warmed in the lower troposphere in association with sensible heat fluxes from the underlying sea surface, contributing to the formation of the cyclone’s warm core.

Differential diabatic heating occurring within regions of deep convection along Invest 91C’s bent-back warm front redistributes PV in the vertical over the cyclone center between 0000 UTC 30 October and 0000 UTC 31 October, eroding upper-tropospheric PV over the lower-tropospheric relative vorticity maximum. The lower-tropospheric relative vorticity maximum separates from the bent-back warm front following the cyclone’s warm seclusion, allowing Invest 91C to completely transition into an axisymmetric, warm-core TC via air–sea interaction processes (e.g., Emanuel 1986) by early November. Figure 19, a cyclone phase space diagram (Hart 2003) summarizing Invest 91C’s life cycle, reveals that the cyclone began to lose its TC

Fig. 20. The (a) track and (b) intensity forecasts of Invest 91C, created from the 11-member GEFS reforecast initialized at 0000 UTC 2 Nov 2006. The thin colored lines in (a) and (b) represent individual ensemble members. The thick black line in (a) and (b) represents the 0.5° NCEP CFSR analysis.
characteristics prior to making landfall along the coast at approximately 1600 UTC 3 November.

c. Eastern North Pacific TT rarity

The TT of Invest 91C was an unprecedented event in the eastern North Pacific, forming in a portion of the basin historically devoid of tropical cyclogenesis (McTaggart-Cowan et al. 2013). According to McTaggart-Cowan et al. (2013), the portion of the eastern North Pacific within which TT occurred would climatologically favor TCs developing in the presence of 1) an upper-tropospheric disturbance and 2) strong lower-tropospheric thermal gradients if tropical cyclogenesis were to take place (their Fig. 4). These characteristics, which are consistent with the strong EC paradigm for TT developed by Davis and Bosart (2004), accurately describe the environment within which Invest 91C formed. The TT of Invest 91C is also noteworthy for occurring in mid-Fall, considerably later in the eastern North Pacific TC season than the majority of other TT events occurring during 1948–2010 (McTaggart-Cowan et al. 2013, their Fig. 9).

The unique aspects of Invest 91C’s life cycle, including its rapid movement toward the west coast of North America between 0000 UTC 2 November and 1800 UTC 3 November (Fig. 2a), likely presented challenges to operational forecasters tasked with its prediction. Figure 20 depicts the variability associated with track and intensity of Invest 91C in the 42-h GEFS reforecast initialized at 0000 UTC 2 November. The majority of the 11 GEFS ensemble members underestimate the eastward movement of the cyclone (Fig. 20a) and overestimate its intensity (Fig. 20b). Indeed, only one ensemble member (shown in gray) accurately captures the position and intensity of Invest 91C in the 0.5° NCEP CFSR dataset (shown in black) at 1800 UTC 3 November. The various diagnostics presented in this study may aid operational forecasters tasked with the future prediction of TT events in the eastern North Pacific if utilized in real time. Additional diagnostics that may be used by operational forecasters to identify and forecast TT events will be the focus of a subsequent study.

It is important to note that Invest 91C, though shown to be a TC in the present study, is not included in the International Best Track Archive for Climate Stewardship (IBTrACS) dataset (Knapp et al. 2010) or in the global climatology of tropical cyclogenesis events constructed by McTaggart-Cowan et al. (2013). The authors suggest that Invest 91C, as well as TCs forming via the TT process in other unusual basins (i.e., the Mediterranean Sea), should be considered for inclusion in these datasets in order to obtain a comprehensive understanding of the frequency and location of tropical cyclogenesis events around the globe.

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