

Diagnostic study of diabatic heating and potential vorticity during a case of cyclogenesis

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सार – 16-17 नवंबर, 2015 को भारी वर्षा के साथ उत्पन्न हुआ चक्रवात सउदी अरब के उत्तरी और मध्य क्षेत्रों से टकराया। इस शोध कार्य में प्रति रुद्धोष्म उष्मण (डाएबैटिक हीटिंग) और विभव भ्रमिल के विश्लेषण के आधार पर इस भारी वर्षा का नैदानिक अध्ययन किया गया है। सिनॉप्टिक विश्लेषण की जाँच परख से यह पता चलता है कि जिनकी वजह से यह घटना हुई उसके गतिकीय कारकों में क्षोभमंडलीय अस्थायित्व से लाल सागर द्रोणी का उत्तराभिमुख विस्तार, अरेबियन प्रायद्वीप में प्रतिचक्रवात, ट्रैवलिंग मध्य अक्षांशीय ऊपरी द्रोणी, आद्रता परिवहन मार्ग और ऊपर की ओर जाने वाली प्रबलगति निहित है। प्रतिरुद्धोष्म उष्मण का परिकलन थर्मो डायनामिक समीकरण के साथ किया गया जिससे यह पता चला कि इस अध्ययन की अवधि के दौरान ऊर्ध्वाधर तापमान एडवेक्शन और एडियोबैटिक परस्पर विपरीत थे। पहले दो दिनों में क्षैतिजीय एडवेक्शन का सबसे अधिक योगदान रहा जबकि विकसित अधिकतम दिनों में क्षैतिजीय उष्मण एडवेक्शन का अधिक योगदान रहा। इस शोधपत्र के अध्ययन में डायनेमिक की जांच भी आइसोबारिकविभव भ्रमिलता के संदर्भ में की गई। इस जांच से यह पाया गया कि निम्न सतही विभव विसंगति का स्थान और विभव भ्रमिलता उत्पन्न होने के स्थान का आकलन उष्मण क्षेत्र से मिलता है जो निम्न सतह विभव भ्रमिल विसंगति के अस्तित्व के पर्याप्त स्रोत के संघनन में सहयोग देता है।

ABSTRACT. On 16-17 November, 2015, north and middle regions of Saudi Arabia were hit by a case of cyclogenesis associated with heavy rainfall. This work presents a diagnostic study of this heavy rainfall case based on the analysis of diabatic heating and potential vorticity. The synoptic analysis investigate that the important dynamical factors that causes this case are the northward extension of Red Sea Trough, anticyclone over the Arabian Peninsula, a travelling midlatitude upper trough, moisture transport pathways and strong upward motion arising from tropospheric instability. The calculation of diabatic heating by the thermodynamic equation illustrate that the contribution of vertical temperature advection and the adiabatic term are opposite to each other during the period of study. The largest contribution of the horizontal cold advection occurs during the first two days while the largest contribution of the horizontal warm advection occurs during the maximum development days. The dynamics of the studied case are also investigated in terms of isobaric Potential Vorticity. It is found that the location of the low-level Potential Vorticity anomaly and the Potential Vorticity generation estimates coincides with the heating region, which implies that condensation supports a large enough source to explain the existence of the low-level Potential Vorticity anomaly.

Key words – Potential vorticity, Diabatic heating, Moisture processes, Heavy rainstorm, Divergent and rotational wind, Red sea trough, Saudi Arabia.

1. Introduction

One of the main and significant force that drives atmospheric circulation is Diabatic heating. The diabatic heating of the atmosphere consists of radiative heating and cooling and release of latent heat and sensible heating. As a result of atmospheric heating cannot be directly measured and must be estimated indirectly from other variables, the understanding of diabatic heating has been hindered. Earlier researches have shown that diabatic potential vorticity (PV) maxima in the lower troposphere

can influence on the onset of extratropical cyclones (Davis and Emanuel, 1991; Davis, 1992; Stoelinga, 1996; Plant *et al.*, 2003).

In numerous cases, the diabatic PV maximum enhances the mutual interaction and coupling between the surface potential temperature (θ) maximum and the upper-tropospheric PV maximum, leading to a powerful feedback process and a more intense cyclone (Davis *et al.*, 1993; Stoelinga, 1996). Contrarily, in other cases, the diabatic PV maximum can obstacle tropopause

and surface waves interaction (Davis and Bosart, 2003). The wind field in lower-tropospheric layer is significantly influenced by the diabatically generated PV maxima in this layer of the atmosphere where there is a lot of moisture. This helps in strengthening the low-level jet that can significantly modify moisture transport and ultimately precipitation distribution (Whitaker *et al.*, 1988; Lackmann and Gyakum, 1999; Brennan and Lackmann, 2005). During boreal autumn, winter and spring, midlatitude cyclonic disturbances can intrude into the Middle East. In combination with the low-level northward advection of warm, moist air from nearby tropical water surfaces, this can lead to a weakened troposphere and rainfall over the Arabian Peninsula (Barth and Steinkohl, 2004; Evans *et al.*, 2004; Chakraborty *et al.*, 2006; Evans and Smith, 2006; Kumar *et al.*, 2015; Mashat and Abdel Basset, 2011). The northward extended low-level trough from the tropical low-pressure over equatorial Africa region across the Red Sea and is usually referred to as the Red Sea Trough or Sudan Low (El Fandi, 1946; Alpert *et al.*, 2004a,b; Tsvieli and Zangvil, 2005). Its formation can be attributed to the interaction between the topography in the Red Sea region and the easterly low-level flow (De Vries *et al.*, 2013). The subtropical anticyclone at the eastern flank of the Red Sea Trough is located over the Arabian Peninsula (AP) and the Arabian Sea at lower- and middle-tropospheric levels and is known as the Arabian anticyclone (Raziei *et al.*, 2012; De Vries *et al.*, 2013). Extreme precipitation events in arid and semiarid subtropical regions often result from tropical–extratropical interactions. The dynamics of such extreme events in subtropical regions, including northwestern Africa, South Africa, Pakistan, the Levant and southwestern North America has been investigated in numerous studies (Dayan *et al.*, 2001; Ziv, 2001; Knippertz *et al.*, 2003; Knippertz, 2005; Knippertz and Martin, 2005, 2007a; Ziv *et al.*, 2005; Hart *et al.*, 2010; De Vries *et al.*, 2013; Martius *et al.*, 2013). Heavy rainfall in central part of AP in winter season is associated with upper- and lower-level cyclonic circulations, potential vorticity (PV) intrusions and enhanced tropospheric moisture content (Kumar *et al.*, 2015). Recently, extreme precipitation and flooding events in the Jeddah region received increased attention in modelling studies (Haggag and El-Badry, 2013; Deng *et al.*, 2015) and diagnostic analyses (Al-Khalaf and Basset, 2013), while De Vries *et al.* (2013) addressed the Jeddah flooding of November 2009 in the context of the Active Red Sea Trough phenomenon. These studies described synoptic and mesoscale conditions: a stationary Arabian anticyclone which favoured moisture transport over the Arabian and Red Seas towards Jeddah, a cold midlatitude upper-level trough that amalgamated with the Red Sea Trough, low-level convergence, upslope winds over the coastal zone and deep moist convection. This work aims to

study the relationship between diabatic processes and the generation of low level-potential vorticity anomaly during a case of interaction between middle latitude cyclone and extratropical cyclones over a limited area during 14 to 19 November, 2015. The behavior of divergent and rotational moisture-flux along with precipitable water content throughout the period of study are also examined.

2. Theoretical aspects

2.1. The diabatic heating

From the elementary concept of thermodynamic we find that

$$\dot{Q} = C_p \frac{dT}{dt} - \alpha \omega \quad (1)$$

where, $\dot{Q} = \frac{dQ}{dt}$ is the diabatic heating rate per unit mass, T the temperature, α the specific volume, C_p the specific heat at constant pressure, $\omega = dP/dt$. By rearranging the terms and expanding dT/dt in Equation (1), the thermodynamic energy equation in isobaric coordinates per unit mass can be written as:

$$C_p \frac{\partial T}{\partial t} = -C_p V \cdot \nabla T - C_p \omega \left(\frac{\partial T}{\partial P} - \frac{\alpha}{C_p} \right) + \dot{Q} \quad (2)$$

where, V is the horizontal wind vector and ∇ the two-dimensional del operator on the isobaric surface. According to Equation (2), $\dot{Q}$ can be estimated as a residual:

$$\dot{Q} = C_p \left[\frac{\partial T}{\partial t} + V \cdot \nabla T + \omega \left(\frac{\partial T}{\partial P} - \frac{\alpha T}{C_p P} \right) \right] \quad (3)$$

Hence, the diabatic heating changes can be determined by quasi-horizontal temperature advection ($V \cdot \nabla T$), adiabatic temperature changes associated with work done on ($\omega > 0$) or by ($\omega < 0$) the air parcel on the environment during vertical displacements ($-\omega \alpha / C_p$), local temperature changes, $(\partial T / \partial t)$ and vertical temperature advection ($\omega \partial T / \partial P$). The release of latent heat due to condensation of turbulent flux, water vapour and short-long wave radiations are the main processes that contribute to.

2.2. Estimation of potential vorticity

Potential vorticity variables were computed from the available atmospheric parameters that are temperature and horizontal wind components on constant pressure surface. Furthermore, the potential vorticity in isobaric coordinates was estimated from the product of potential temperature

gradients and the vertical components of absolute vorticity as:

$$PV = \left[\left(\frac{\partial v}{\partial x} - \frac{\partial u}{\partial y} \right) + f + \frac{R}{\sigma P} \left(\frac{\partial v}{\partial p} \frac{\partial T}{\partial x} - \frac{\partial u}{\partial p} \frac{\partial T}{\partial y} \right) \right]_p \frac{\partial \theta}{\partial p} \quad (4)$$

where, f is the Coriolis parameter, u the west-east (zonal) wind component and v the south-north (meridional) wind component and θ the potential temperature. According to WMO (1986), the dynamic tropopause can be defined as the potential vorticity with $P = 1.6 \times 10^{-7} \text{ Kpa}^{-1} \text{ s}^{-1} = 1.6 \text{ PVU}$, where, for convenience, the potential vorticity unit (PVU) is set as $10^{-7} \text{ Kpa}^{-1} \text{ s}^{-1}$.

2.3. Moisture flux calculation

According to Chakraborty *et al.*, 2006, the precipitable water content in an air column is calculated as:

$$\text{PWC} = \frac{1}{g} \int_{P_u}^{P_L} q \, dp \quad (5)$$

Similarly, zonal and meridional water vapour flux transports are calculated as:

$$Q_u = \frac{1}{g} \int_{P_u}^{P_L} q u \, dp \quad (6)$$

and

$$Q_v = \frac{1}{g} \int_{P_u}^{P_L} q v \, dp \quad (7)$$

where, total water vapour transport is $Q = Q_u i + Q_v j$.

The total water vapour flux decompose into rotational and divergent components (Rosen *et al.*, 1979; chen, 1985) in terms of the stream function (ψ) and velocity potential (χ), we can write

$$Q = kx \nabla \psi + \nabla \chi \quad (8)$$

The ψ and χ fields are obtained from

$$\nabla^2 \psi = k \cdot \nabla \chi \quad \text{and} \quad \nabla^2 \chi = \nabla \cdot Q \quad (9)$$

The Equation (9) can be solved easily for ψ and χ using the relaxation method (Krishnamurti and

Bounoua, 1996). Finally, rotational and divergent moisture vectors can be derived as:

$$Q_u^R = -\frac{\partial \psi}{\partial y}, Q_v^R = \frac{\partial \psi}{\partial x} \quad \text{and} \quad Q_u^D = \frac{\partial \chi}{\partial x}, Q_v^D = \frac{\partial \chi}{\partial y} \quad (10)$$

where, g acceleration due to gravity; q refers to the specific humidity; P_L pressure at the bottom of the air column; and P_u pressure at the top of the air column.

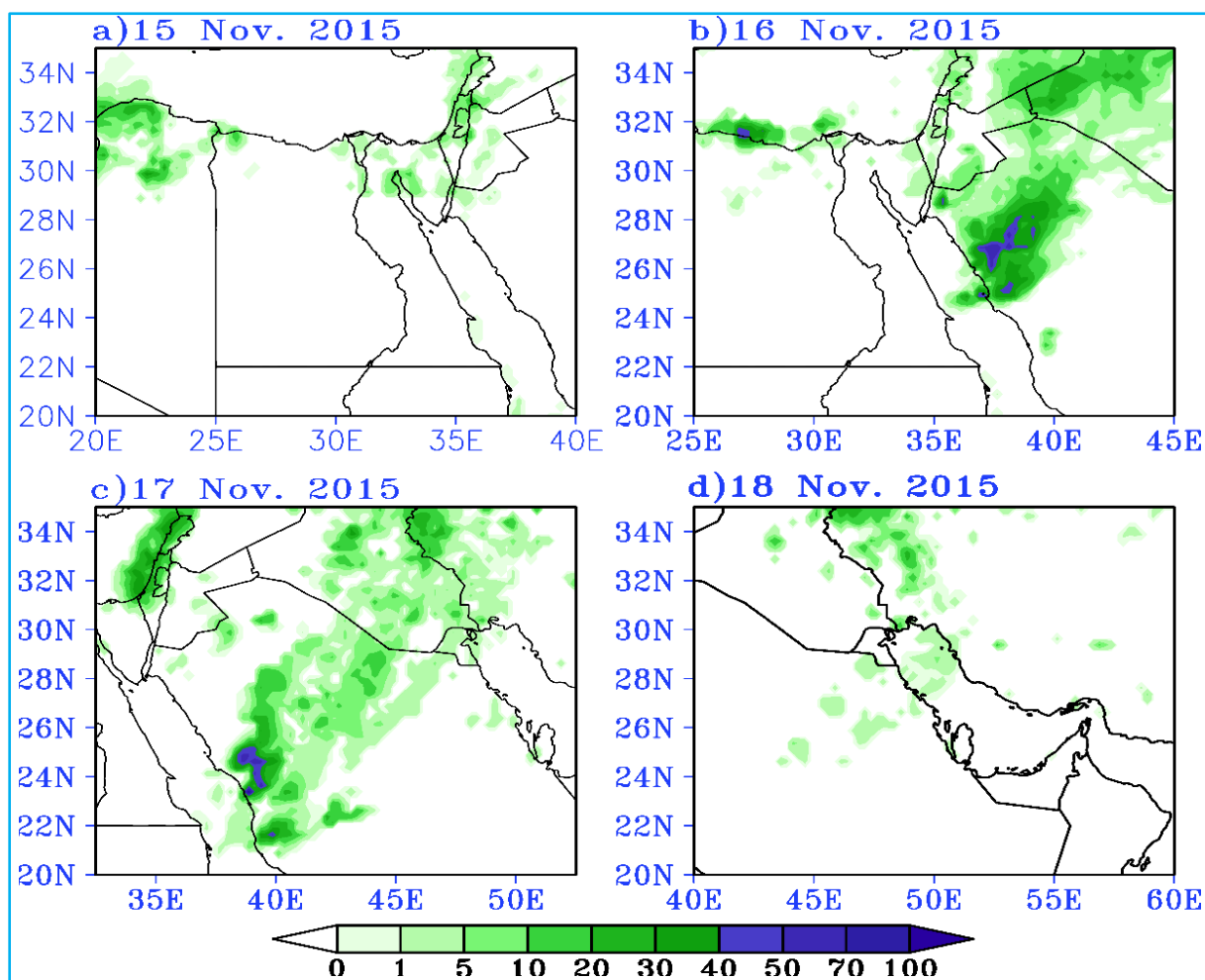
3. Data and computation

3.1. Data

Six-hourly of zonal (u) and meridional (v) wind components (m/sec), temperature (T °C), specific humidity (q g/kg) and the geopotential height (Z gpm) with horizontal resolution of $2.5^\circ \times 2.5^\circ$ over the area bounded by 20° W - 80° E and 0° - 80° N from the European Centre for Medium-range Weather Forecasts Reanalysis Interim dataset (Dee *et al.*, 2011) during the period from 13 to 20 November, 2015. These parameters are obtained at 1000, 850, 700, 500, 400, 300, 250, 200, 150 and 100 hPa pressure levels. Daily precipitation data of $0.25^\circ \times 0.25^\circ$ horizontal resolution were obtained from Tropical Rainfall Measuring Mission (TRMM-3B42) (Huffman *et al.*, 2010).

3.2. Analytical procedures

The diabatic heating is calculated using Equation 3 at 0000, 0600, 1200 and 1800 UTC. Therefore, time derivatives evaluated by centered finite difference spanning 12 hours gives a significant indication of the time variation of heating. The centered finite differences are used to compute both horizontal and vertical derivatives, except those at 1000 and 100 hPa where non-centered differences were employed. The inner domain used for calculating the terms of the thermodynamic Equation (3) extends from 30° E to 50° E and from 15° N to 40° N . Each term in the right-hand side of Equation (3) is diagnostically estimated, except the tendency term is computed using a centered finite differences scheme. The diabatic heating term $\dot{Q}$ has been determined as a residual, by summing all terms in the right-hand side of Equation 3. The diabatic heating, represented in terms of the thermodynamic equation in isobaric coordinates, is affected by the evaluation of the vertical motion (ω). This motion is calculated using the Q-vector representation of the quasi-geostrophic ω equation (Bluestein, 1992), using the relaxation method (Krishnamurti and Bounoua, 1996). All terms in Equation (3) are expressed as heating rates per unit mass and can be converted to daily heating by the following transformation, $1 \text{ W Kg}^{-1} = 86 \text{ K d}^{-1}$. The centered finite differences were also used to calculate the horizontal



Figs. 1(a-d). The spatial rainfall pattern of TRMM data for the period 15- 18 November, 2015

and vertical derivatives of the potential vorticity Equation (4) except those at 1000 and 100 hPa.

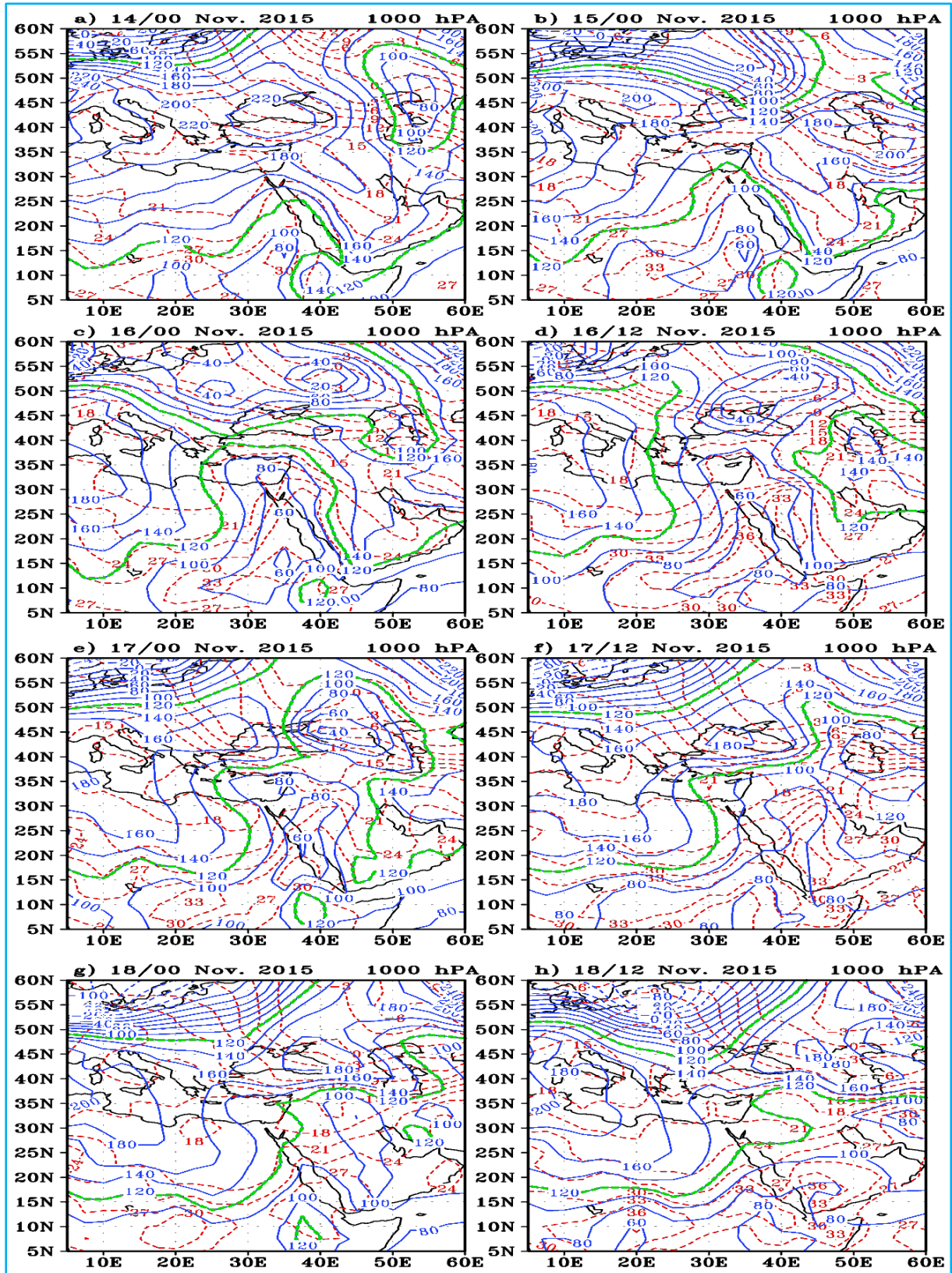
The precipitable water content in an air column within any two pressure layers was calculated using equation 5. To determine the moisture source (divergence) and moisture-sink (convergence) regions of water vapor, we calculate both the divergent and rotational components of moisture transport. To find the components of moisture vectors, first we calculated the total water vapor transport (Q) within any two pressure layers using equations (6) and (7), then we used equation (9) for stream function (ψ) and velocity potential (χ). Finally, we used equation (10) because the stream function and velocity potential are already known.

4. Synoptic discussion

Our case study represents a common case of cyclogenesis over east Mediterranean. It has long been

recognized that cyclogenesis occurs typically during the interaction between upper air middle latitude cyclone with the Red Sea Trough (RST) which represents the northward extension of the Sudan monsoon low over the east Mediterranean. On 16, 17 and 18 November, 2015, heavy rainstorms hit northwest, middle and northeast regions Saudi Arabia respectively. The rainfall recorded by the meteorological stations in Tabuk, Wajah and Maka were 48, 46 and 25 mm respectively during the three days. Figs. 1(a-d) presents the pattern of rainfall from the TRMM data for four times in the period of study. Geopotential height in meters (gpm) and temperature in °C on the 1000 hPa and 500 hPa charts at 0000 and 1200 UTC on each day of the period 14-18 November are shown in Figs. 2(a-h) & 3(a-h), respectively.

At 0000 UTC 14 November, Fig. 2(a) shows that the subtropical high pressure dominates over North Africa and the Mediterranean area; it also extends easterly to cover the north part of Saudi Arabia and the east Mediterranean



Figs. 2(a-h). 1000 hPa height contour in 20 m intervals (solid) and temperature (dotted) in 5 °C increments for 14/00 - 18/12 UTC November, 2015

countries. Fig. 2(a) also illustrates that the Sudan low and its associated inverted V-shaped trough (RST) oscillate northward to cover the eastern part of Egypt and the entire Red Sea region. An obvious thermal gradient coupled with the northward oscillation of RST extends zonally to cover the southern part of Saudi Arabia and the northern part of Sudan. Through the next 24 hours (1200 UTC 15 November), the subtropical high decays and moves eastward, whereas the RST propagates slowly northward and the Sudan low deepens to 80 gpm [Fig. 2(b)]. On 500 hPa, the upper air system first appears as an extension of the traveling depression, west of Libya at 0000 UTC 14 November [Fig. 3(a)]. The cut-off low is deepening and moved eastward at 0000 UTC 15 November to become clear over east of Egypt [Fig. 2(b)]. At 0000 UTC 15 November, a thermal gradient lies along the northwest of Saudi Arabia; this baroclinic zone at the lower layer helps in the development of an upper level cut-off low. At 0000 UTC 16 November an intense development occurs at the surface and in the upper atmospheric layers, where the surface Sudan low and its associated RST moves northward and cover the northern region of the East Mediterranean, Egypt and Saudi Arabia.

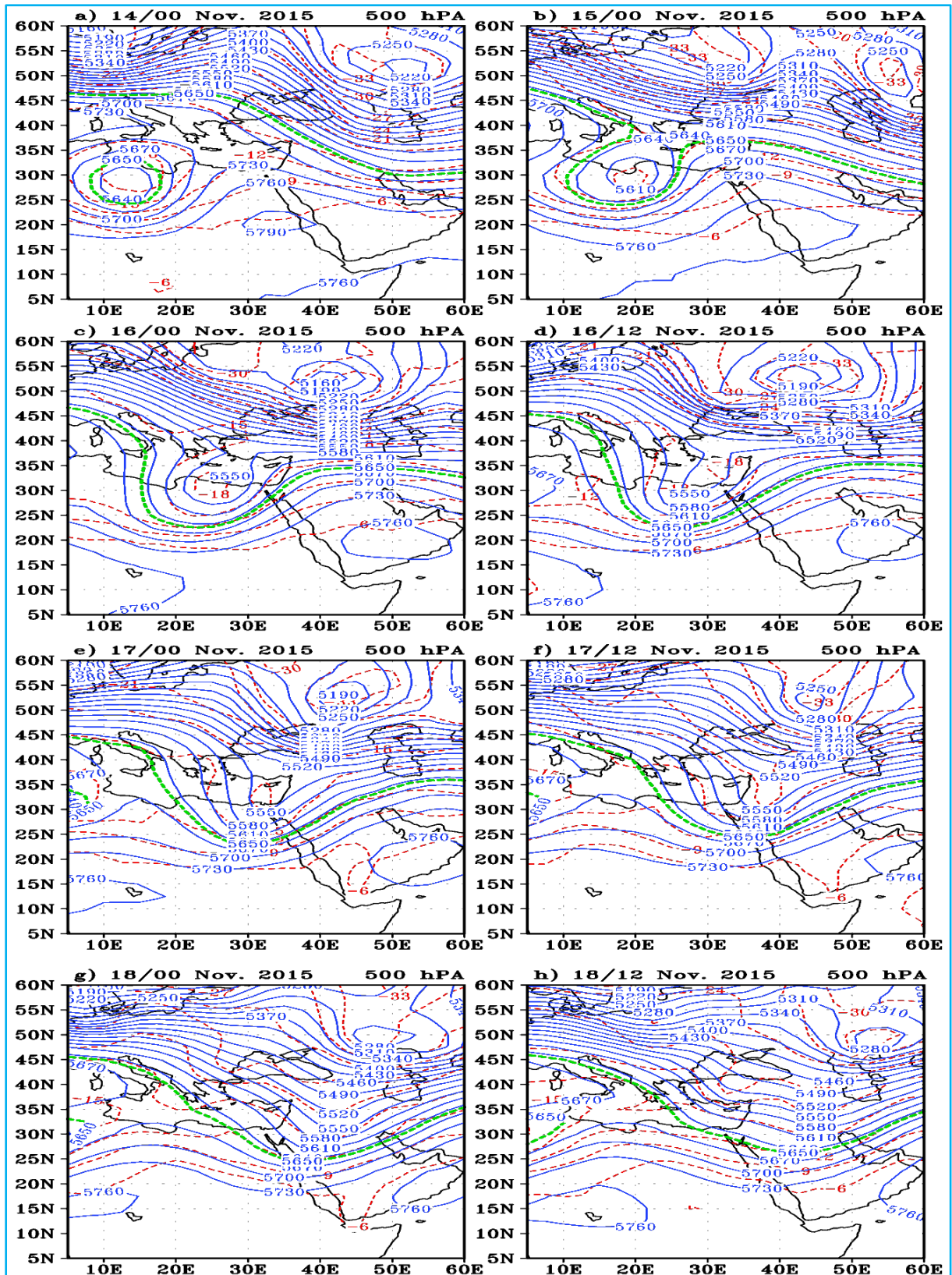
The Sudan low is cantered over the Red Sea, while the geopotential height at its center reaches 60 gpm. On the other hand, the cut-off low in the upper atmosphere (500 hPa) deepens and moves slowly eastward to reach just north of Egypt; the geopotential height at the center reaches 5550 gpm [Fig. 2(c)]. During 1200 UTC 16 November, the inverted V-shaped trough (RST) associated with the Sudan low oscillates northward. On 500 hPa, the cut-off low also moves eastward to reach just northeast of Egypt. During the period 1800 UTC 16 November to 0600 UTC 18 November (the rainy days), a strong interaction occurs between the inverted V-shaped troughs extending from the tropical region and from mid-latitude region. These two depressions amalgamate to develop a single system. Their most important features are the strong tropical northward warm advection associated with the air flow around the Sudan low, as well as the polar strong southward cold advection. The interaction between these two (tropical and polar) air masses leads to a great deal of atmospheric instability over the East Mediterranean and the west of Saudi Arabia. After 1200 UTC 18 November, the inverted V-shaped trough of the Sudan low moves south-westward while the upper air moves northeastward where the associated cut-off low disappeared and the interaction between these two troughs vanishes. In the next day (19 November), the subtropical high pressure over North Africa and the western Mediterranean is extended with a major ridge that joins the Siberian high.

5. Diabatic heating and moisture transport study

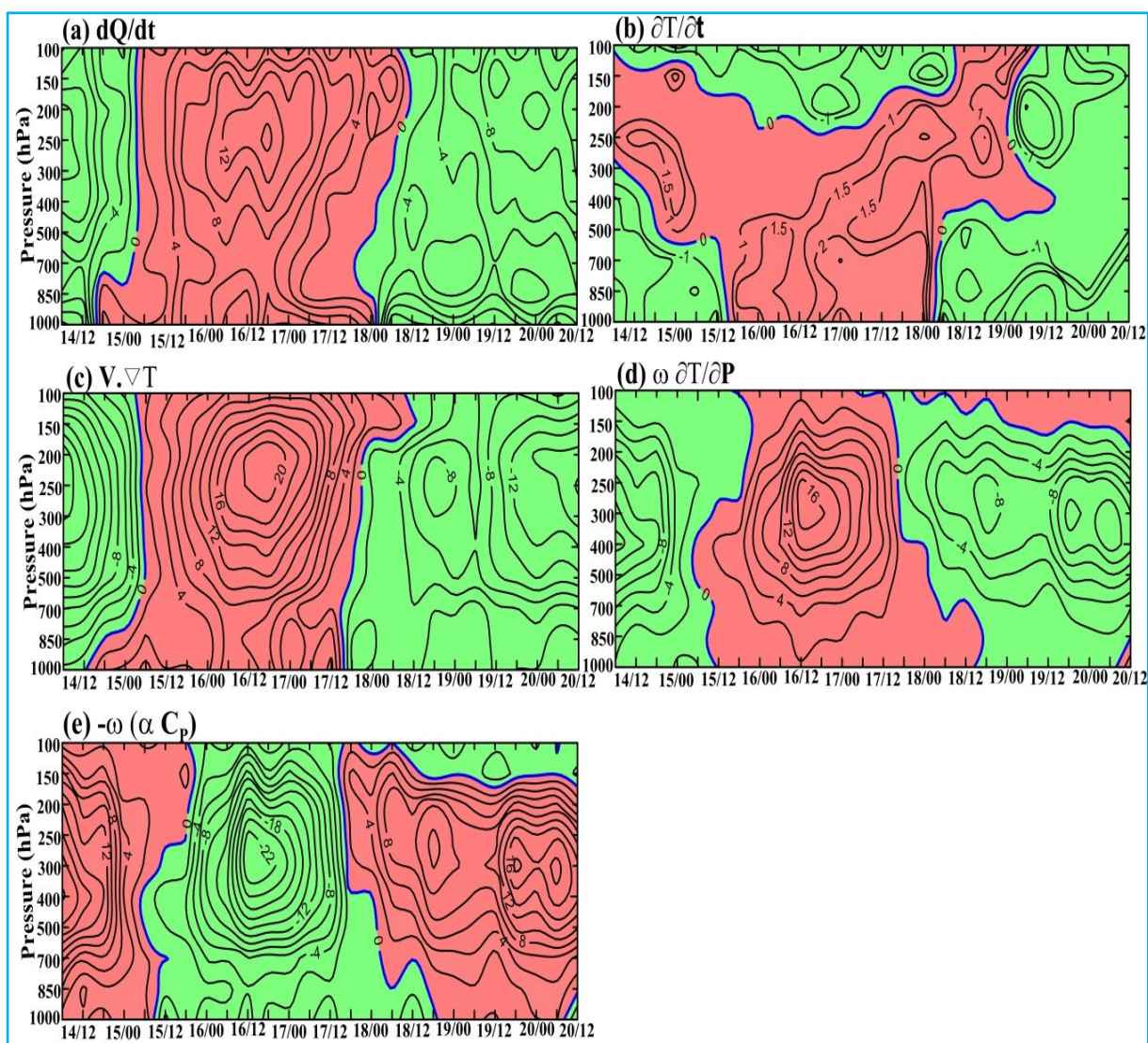
The assessment of the diabatic heating in the isobaric thermodynamic method terms allows the individual contribution of each effect to be investigated (Budyko, 1974). Nevertheless, using the thermodynamic equation in isobaric coordinates provides a way to explain how the diabatic heating is balanced by the sensible heat horizontal advection and adiabatic heating as a result of vertical motion. The inner domain used for calculating the terms of the thermodynamic Equation 3 extends from 30°-50° E and from 15°-40° N.

5.1. Time-pressure variations of the thermodynamic equation terms

Figs. 4(a-e) shows the time height cross section of the values of each term in the thermodynamic Equation 3 throughout the period of study. Fig. 4(b) shows that the contributions from the term of local temperature changes to the diabatic heating changes are very small comparing with all the other terms. It illustrates that there is a decrease of temperature below 400 hPa during the period 14/00 to 15/12. At the rainy days (16-18 November) there is a local increase of temperature from the surface up to 200 hPa while above 200 hPa there is a local decrease of temperature. A local decrease of temperature occurs in the period 19/00 to 20/12 November at all levels. It is evident that the largest contribution of the horizontal cold advection ($V \cdot \nabla T$) of temperature exists during the period from 14/00 to 15/00 November at all levels and its maximum values occurs between 500-250 hPa [Fig. 4(c)]. Horizontal warm advection occurs at all pressure levels during the period from 15/12 to 17/18 November with maximum between 700-200 hPa. A considerable horizontal cold advection appears at all levels during the last three days. In the vertically averaged sense, the adiabatic term ($-\omega \alpha / CP$) and vertical advection ($\omega \partial T / \partial P$) work in an opposite sense which reveal that these two terms numerically cancel each other when viewed in the time-pressure plane. Furthermore, these two terms do not balance each other exactly. Figs. 4(d&e) show that the contribution of vertical temperature advection ($\omega \partial T / \partial P$) and diabatic temperature term are opposite of each other. Meaning of the times (areas) of cooling in Fig. 4(d) corresponds with the times (areas) of heating in Fig. 4(e) and *vice versa* which indicate that there is a strong negative correlation between the patterns of the adiabatic temperature term and the vertical advection of temperature term. Fig. 4(d) shows that the vertical advection term ($\omega \partial T / \partial P$) tends to decrease the temperature throughout the period 14/00 to 15/12 November. The maximum heat contribution (increasing of temperature) of this term occurs during the period 15/18 to 17/18 November between 700-200 hPa. A pronounced decrease of



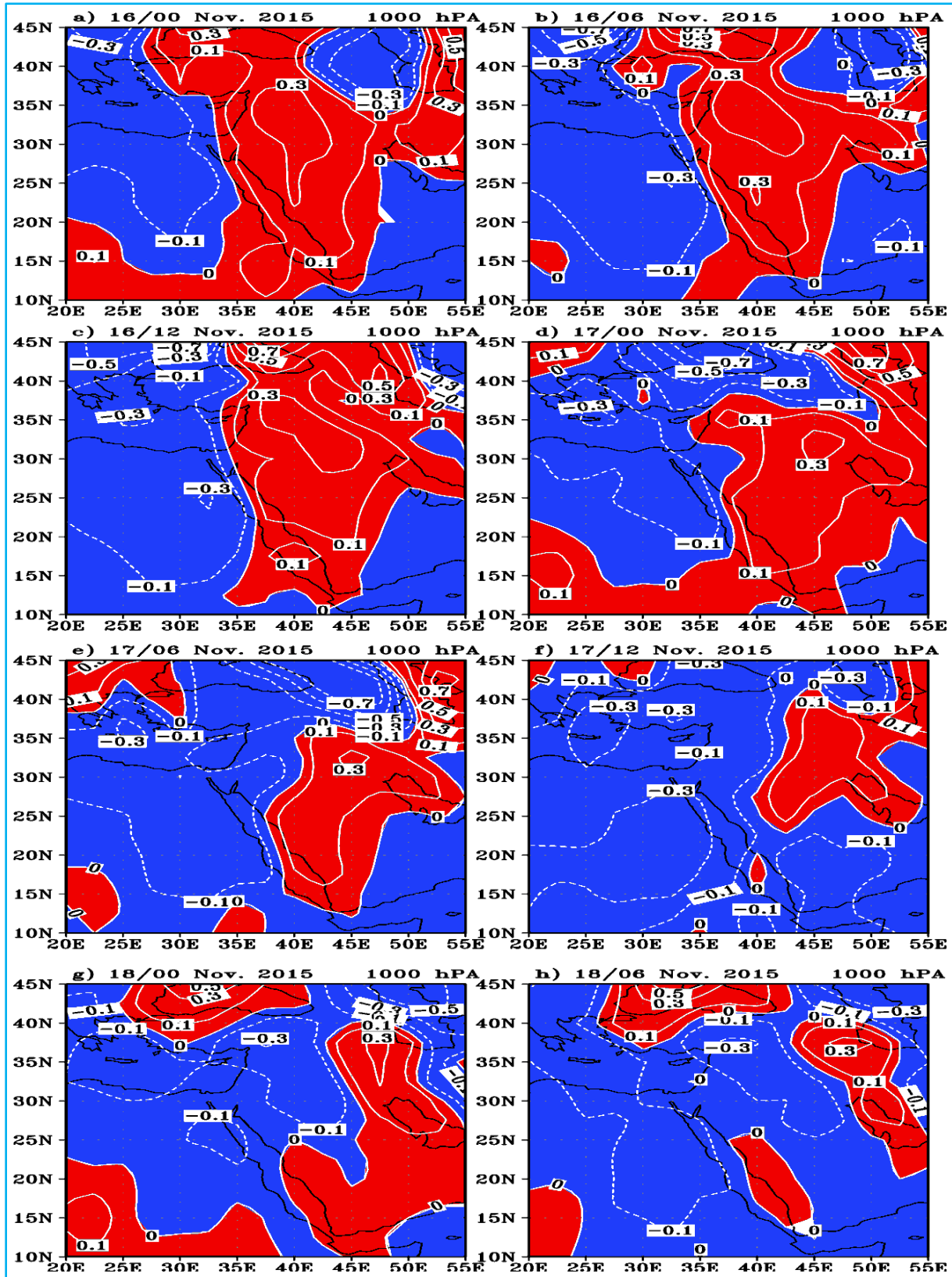
Figs. 3(a-h). 700 hPa height contours in 20 m intervals (solid) and temperatures (dashed) in 5 °C increments for 14/00 - 18/12 UTC November, 2015



Figs. 4(a-e). Time height cross section of (a) diabatic heating, (b) local temperature change, (c) horizontal advection of temperature, (d) vertical temperature advection and (e) adiabatic heating. Unit : 10^{-2} K d^{-1}

temperature by the vertical advection term occurs during the period from 18/00 to 20/12 November specially between 700-200 hPa levels. Fig. 4(e) shows that the adiabatic term ($-\omega\alpha/C_p$) tends to increase the temperature throughout the period from 14/00 to 15/12 November specially between 500-200 hPa. The maximum decrease of temperature by this term occurs between 15/12 to 17/18 November specially between 500-200 hPa. During the last period (18/00 to 20/12) this term increase temperature from the surface to 150 hPa. Fig. 4(a) illustrates the diabatic heating as the sum of the four terms in the right-hand side of equation (3). The outstanding features here are the cooling at all levels during the period 14/00 to 15/00 November with maximum on 14/00 November between 500-150 hPa. A pronounced heating exists during

the period 15/06 to 18/00 November at all levels, its maximum appears above 500 hPa. From 18/00 November to the end of study period a considerable cooling appears at all levels. In the last period the diabatic heating term is affected by the horizontal advection term more than the other terms, where the cooling occurs at all levels and increasing gradually from 19/12 to 20/12 November above 500 hPa. The existence of the Red sea and the Red sea mountains supports the suggestion that the diabatic heating in the lower layer is primarily due to latent heat release on 15/00-17/12 November. The cyclogenetic mechanism in the eastern Mediterranean is firstly due to surface processes and the subtropical jetstream (Alpert and Warner, 1986). Secondly, the major horizontal gradient of advection over the western slopes



Figs. 5(a-h). Horizontal distribution of diabatic heating rate, contour interval is 1° K/day, positive value (red color) negative values (blue color), for 1200 UTC 16-18 November, 2015

of the Turkish mountains probably indicate that the cold northerly air turns around the mountains (EL-Fandy, 1946).

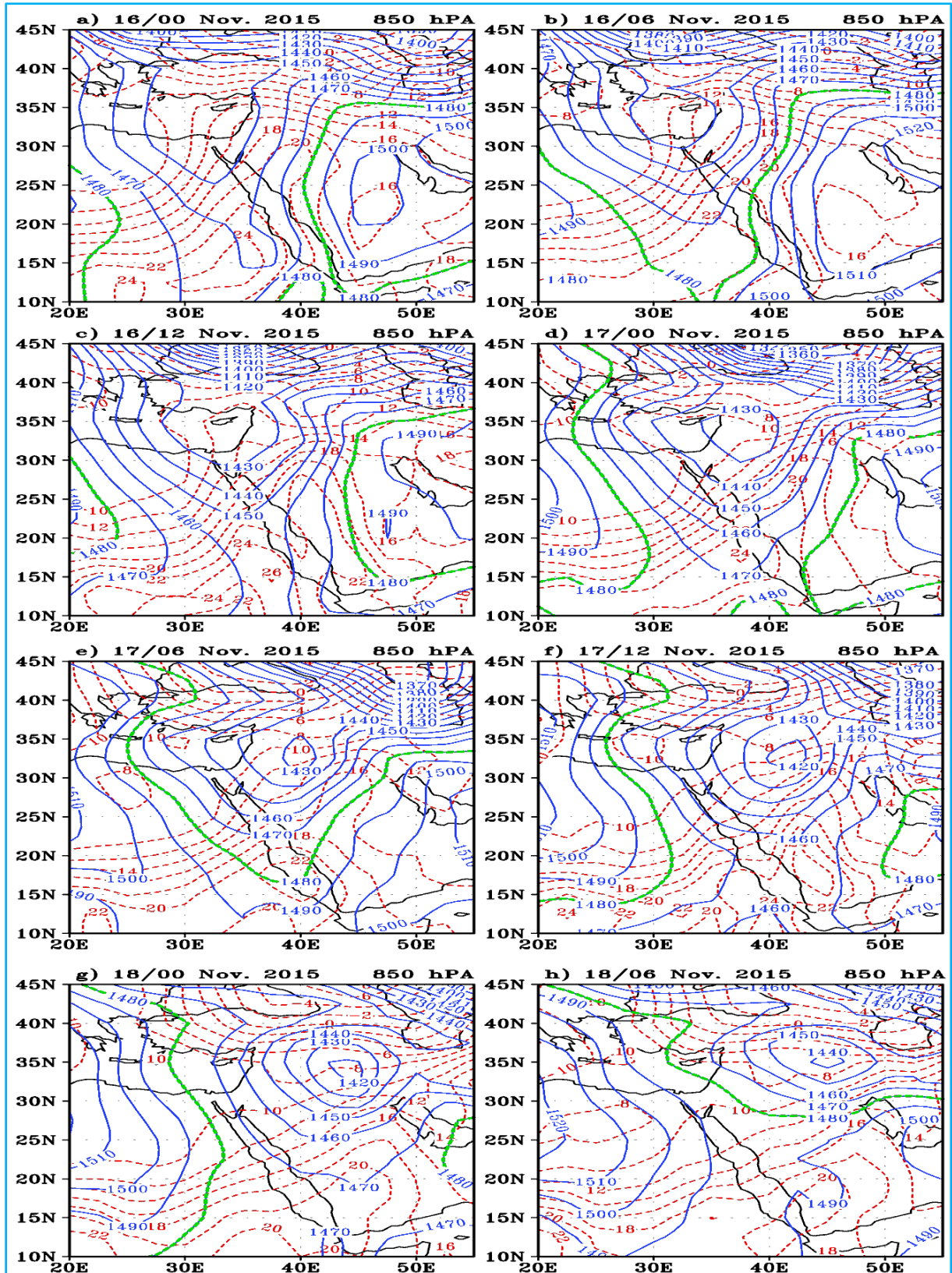
5.2. Spatial distribution of diabatic heating

The diabatic heating rate during the study period from 0000 UTC 14 November to 0000 UTC 19 (14/00 to 19/00) November has been calculated. Figs. 5(a-h) illustrates the diabatic heating rate spatial distribution at 1000 hPa in 8-time steps only starting from 16/00 to 18/06 November. The pattern of diabatic heating in Figs. 5(a-h), shows that the cold sector with diabatic cooling exists at the west of the 1000 hPa trough (approximately west of the Red Sea), while east of the trough the diabatic heating prevails. By comparing the diabatic heating horizontal distribution at 1000 hPa in Figs. 5(a-h) with geopotential height and temperature at 1000 and 850 hPa in Figs. 2(a-h)&6(a-h) respectively, it is indicated that at the western part of the trough (area of downward motion) there is a region of diabatic cooling and at the eastern side of the trough (area of upward motion) there is a dominant region of diabatic heating. In other expressions, the diabatic heating acts as warm advection and the diabatic cooling acts as cold advection. Therefore, diabatic heating (cooling) in the vicinity of developing convection is usually associated with rising (sinking) motions. By looking at the heating and cooling time variation over the inner domain [Figs. 5(a-h)], which is used for the present calculation, two main features of the horizontal distribution of heating rates at 16/00 November can be seen. One feature that stands out clearly is a cooling maximum off the west coast of the Red Sea over Egypt and north of Sudan (the maximum cooling is 0.3 K/6h). The maximum cooling tends to decreasing southward over the southwestern parts of the Red Sea, which implies that the source of cooling is caused by the advection of cold arctic air across most of the Mediterranean Sea. Another notable feature is the temporal variation of the heating pattern over Saudi Arabia. Actually, the temporal variations of these two features are significant. It was detected that the cooling, found over the north-western parts of the Red Sea at 06/00 November, moved slowly eastward between 16/00 to 17/00 November. It was eventually increased, before the center of the heating moved northeastward and was weakened. The cooling over the eastern parts of the Red Sea moved and extended eastward with intensification over northeast of Mediterranean during the period 17/00 to 17/06 November. At the same time, the heating area has shrunk over north-eastern Saudi Arabia and then moved north-eastward to reach over northeast of Arabian Gulf at 18/00 and 18/06 November. Finally, it is interesting to note that the maximum heating (occurring during 16/00 to 17/00

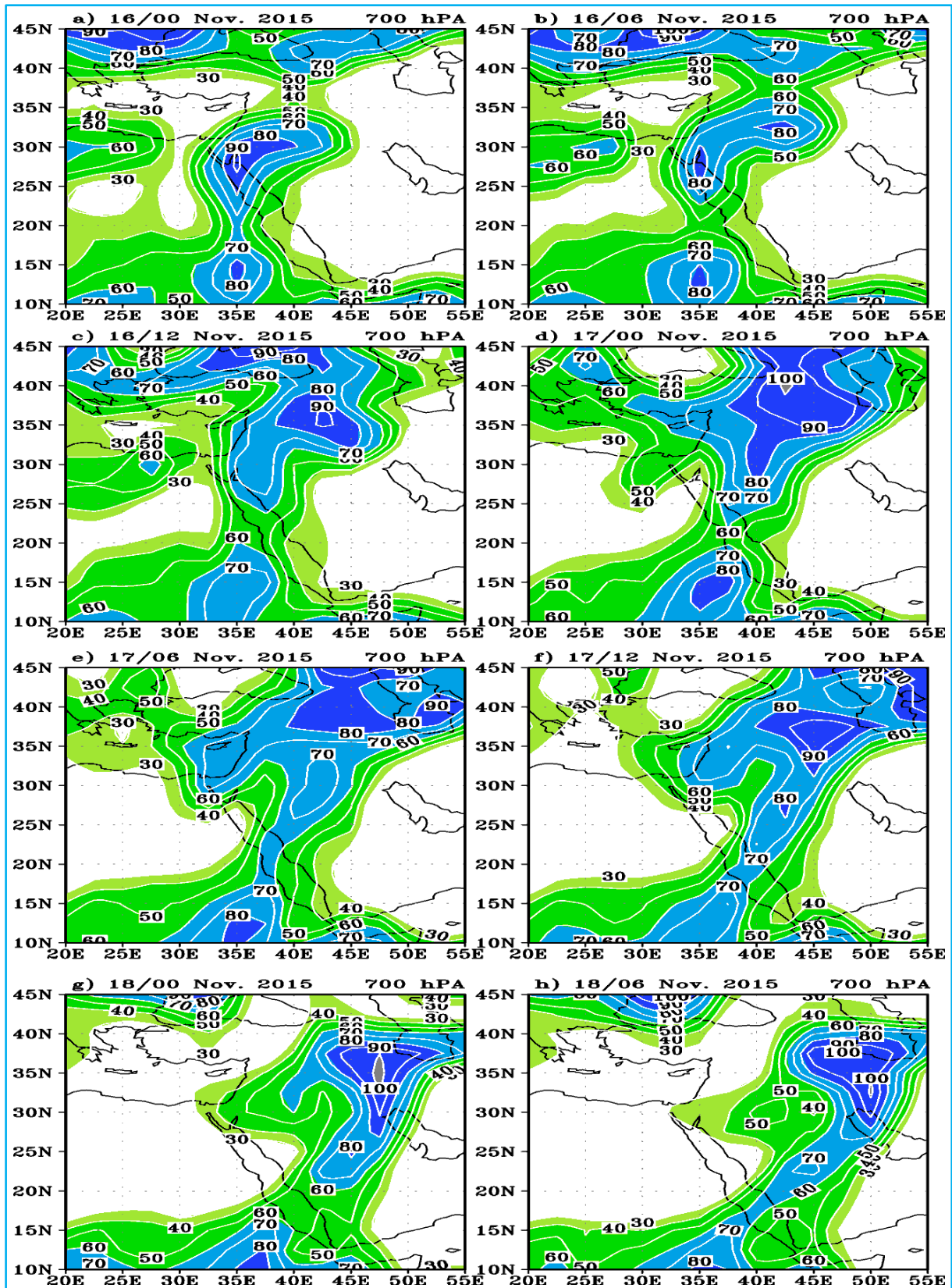
November) was found to be associated with the maximum development of the storm system and also, the movement of the heating or cooling pattern is strongly associated with the movement of the storm.

5.3. Moisture transport

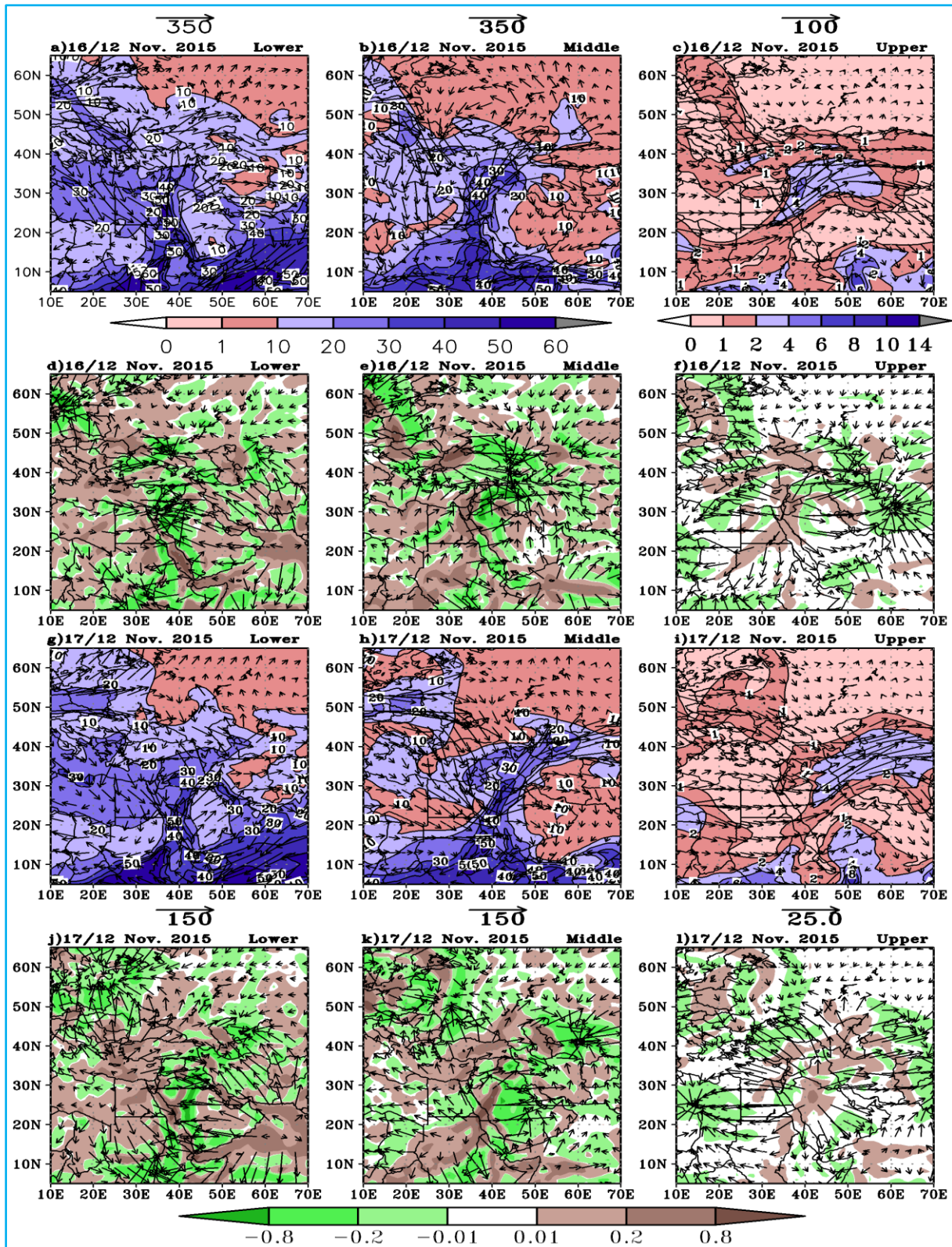
The interaction between the midlatitude and tropical systems supported the transport of tropical moisture towards the Arabian Peninsula (AP). During 16-18 November, enhanced moisture is observed over Arabian and Red Seas, equatorial and north-eastern Africa and AP, leading to above-normal relative humidity over the larger part of northern Saudi Arabia on 16-17 November [Figs. 7(a-h)]. In this section we will illustrate the behaviour of the divergent and rotational components of the total water vapour flux along with precipitable water content for three tropospheric layers at 16 and 17 November. The lower tropospheric layer is taken from 1000 to 850 hPa, while the middle layer from 700 to 500 hPa and upper layer from 400-100 hPa (Chakraborty *et al.*, 2006). Figs. 8(a-l) illustrate the divergent and rotational components of integrated moisture-flux vector along with precipitable water in the three tropospheric layers at 16 and 17 November 2015. The divergent component indicates regions of moisture source or sink, while the rotational component describes the atmospheric water vapor transport. In the lower tropospheric layer, it is clear that there are three moisture-source regions observed from the divergent component [Figs. 8(d&j)]; the first one is located over the Arabian Sea (including south Red sea, north of Ethiopia and middle of Sudan); the second region locates over Russia, while the third region (not shown) locates over the north of Atlantic. These source regions supply moisture to the area of interaction (east of Mediterranean and north of KSA). The presence of a strong anticyclonic circulation associated with the Azores high over north Africa helping on transporting the moisture to the region of interaction [Figs. 8(a&g)]. The rotational moisture transport component brings moisture from two regions; the first region appears over the Indian Ocean, Arabian Sea, south of Red Sea and north east of Sudan. The second source region is the Atlantic and Mediterranean Sea. In the middle layer [Figs. 8(e&k)], the divergent moisture flux area at 16 November is observed over east Mediterranean countries (Syria, Iraq, east of turkey), while at 17 November the areas of divergent moisture flux appear over east and middle east of Europe (its center at 50° N, 30° E). In this layer, the rotational moisture transport component brings a huge amount of moisture with an amount of precipitable water more than 40 mm from two regions; the first source region appears over the Indian Ocean, Arabian Sea, south of Red Sea and north east of Sudan, while the second source region is the



Figs. 6(a-h). 850 hPa height contour in 10 m intervals (solid) and temperature (dotted) in 2 °C increments for 16/00 - 18/06 UTC November, 2015



Figs. 7(a-h). Relative humidity (>30%) at 700 hPa, contour interval is 10% for 16/00 - 18/06 UTC November, 2015



Figs. 8(a-l). Divergent component (second and fourth row) and rotational component (first and third row) of integrated moisture-flux vector ($\text{kg m}^{-1} \text{s}^{-1}$) along with precipitable water in different tropospheric layers; the colored regions in the second and fourth row are the divergent (source) and convergent (sink) component of moisture flux ($10^{-3} \text{ kg m}^{-2} \text{s}^{-1}$) at the same integrated layers for 1200 UTC of 16 and 17 November, 2015

Atlantic and Mediterranean Sea [Figs. 8(b&h)]. In the upper layer [Figs. 8(f&l)], the divergent moisture flux at 16 November is observed over Iran, while at 17 November the areas of divergent moisture flux appear over south of AP, above and north of Turkey and over Libya. Two elongated convergence zones are seen over the area of interaction, the first one [Fig. 8(f)] appears over north of Red sea and north of KSA while the second one [Fig. 8(l)] extended from north of Iran to the middle of Saudi Arabia. In this layer, the rotational moisture transport component brings a very little amount of moisture from two regions; the first source region is from the north Atlantic through south Europe, while the second source associated with the subtropical jet [Figs. 8(c&i)].

The lower and middle tropospheric moisture divergence and the upper tropospheric moisture divergence over east of Mediterranean are indication of the vertical motion of moist air, which in turn may release latent heat due to condensation. This available heat energy may be the source of low-level baroclinic instability. As the cyclone shifts slowly eastward, a divergent moisture flux area is observed over east Mediterranean region at lower tropospheric layer (17 November), while over the same region in the middle tropospheric layer there is a convergent moisture flux which indicates downward motion [Fig. 8(e&k)]. However, the strong divergence area of available moisture over north and middle of Saudi Arabia in the lower tropospheric layer corresponding to strong convergence over the same area in the upper layer may explain the heavy rainfall during 16 and 17 November 2015 [Figs. 1(a-d)]. The upper-tropospheric convergence over north east of Saudi Arabia and Iraq associated with the upper level moisture transport may have a considerable influence on precipitation over this region at 17 and 18 November, 2015.

6. Generation of the low-level PV anomaly

Manabe, 1956 and Kuo *et al.*, 1991 have been observed that there is a strong low-level Ertel's potential vorticity (EPV) anomalies in association with cyclone development. In their cases, the air inside the low-level EPV anomaly was not of stratospheric origin and the EPV anomalies were probably generated by condensation. Figs. 9(a-h) illustrates the position and the development of the low-level EPV anomaly associated with the studied case, while Figs. 7(a-h) demonstrates the relative humidity (RH) pattern for the period from 16/00 to 17/06 November 2015. Where, the low-level EPV anomaly regions are usually associated with high values of RH, which reach about 90%, indicating that the air inside the central region of the EPV anomaly is nearly saturated. Figs. 10(a-h) illustrates the vertical motion at 850 hPa for the period from 16/00 to 18/06 November. It shows the movement of

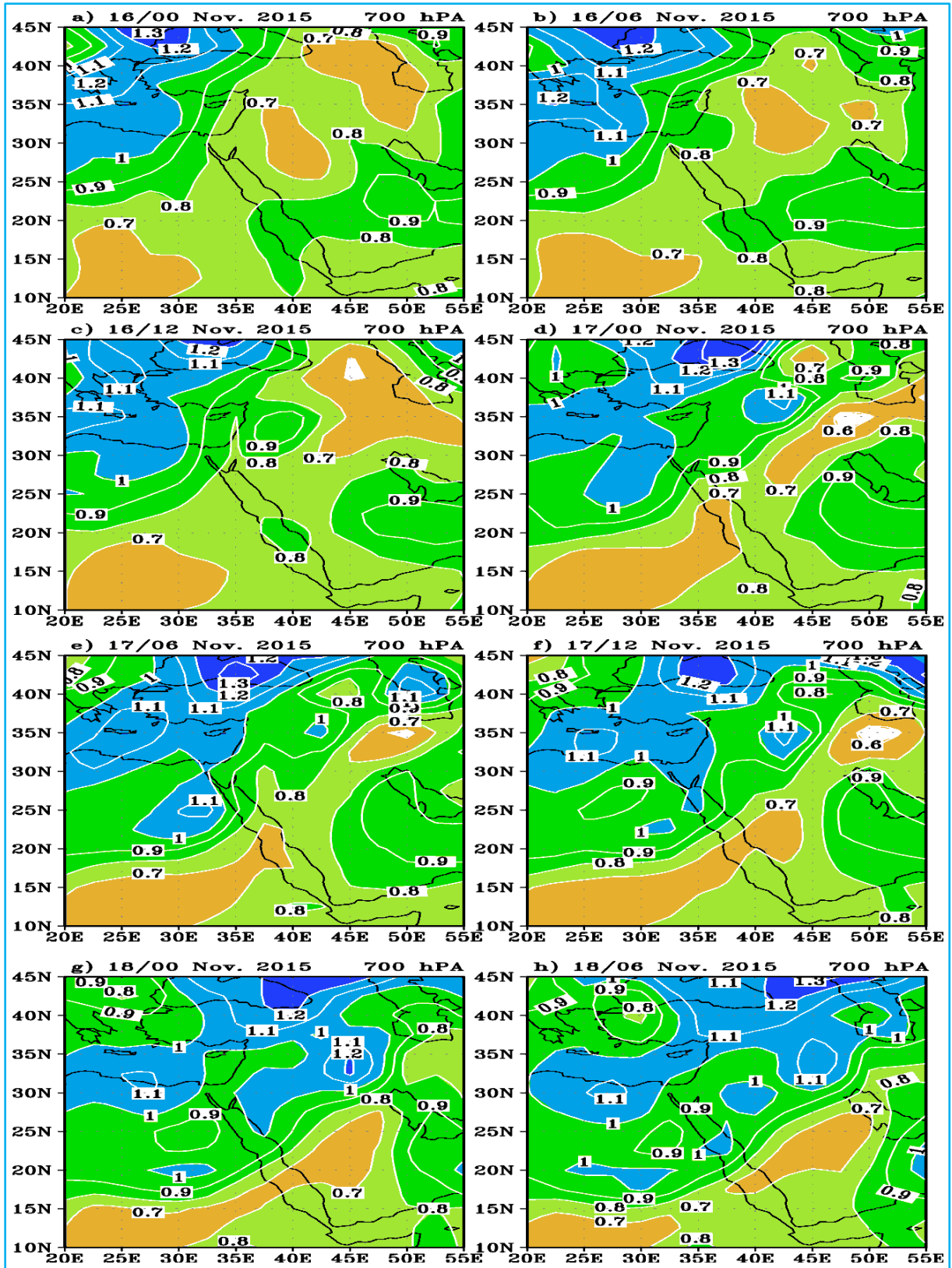
the centers for upward and downward motions with the movement of the storm during its development. Generally, it is clear that the regions of low-level EPV anomaly [Fig. 9(a-h)] are accompanied by the areas of negative vertical motion [Fig. 10(a-h)]; this is verified throughout the period of study. Figs. 5(a-h) & 9(a-h) and the above discussion reveal that the strong diabatic processes occur in the vicinity of the low EPV anomaly. For a more understanding, the estimation of EPV generation rate from latent heat release is performed following the technique used by Davis and Emanuel (1991). The EPV generation form can be written as:

$$\left(\frac{DPV}{Dt}\right)_{LH} = \frac{gk\pi}{p} \eta \cdot \nabla \left(\frac{d\theta}{dt}\right)_{LH} \quad (11)$$

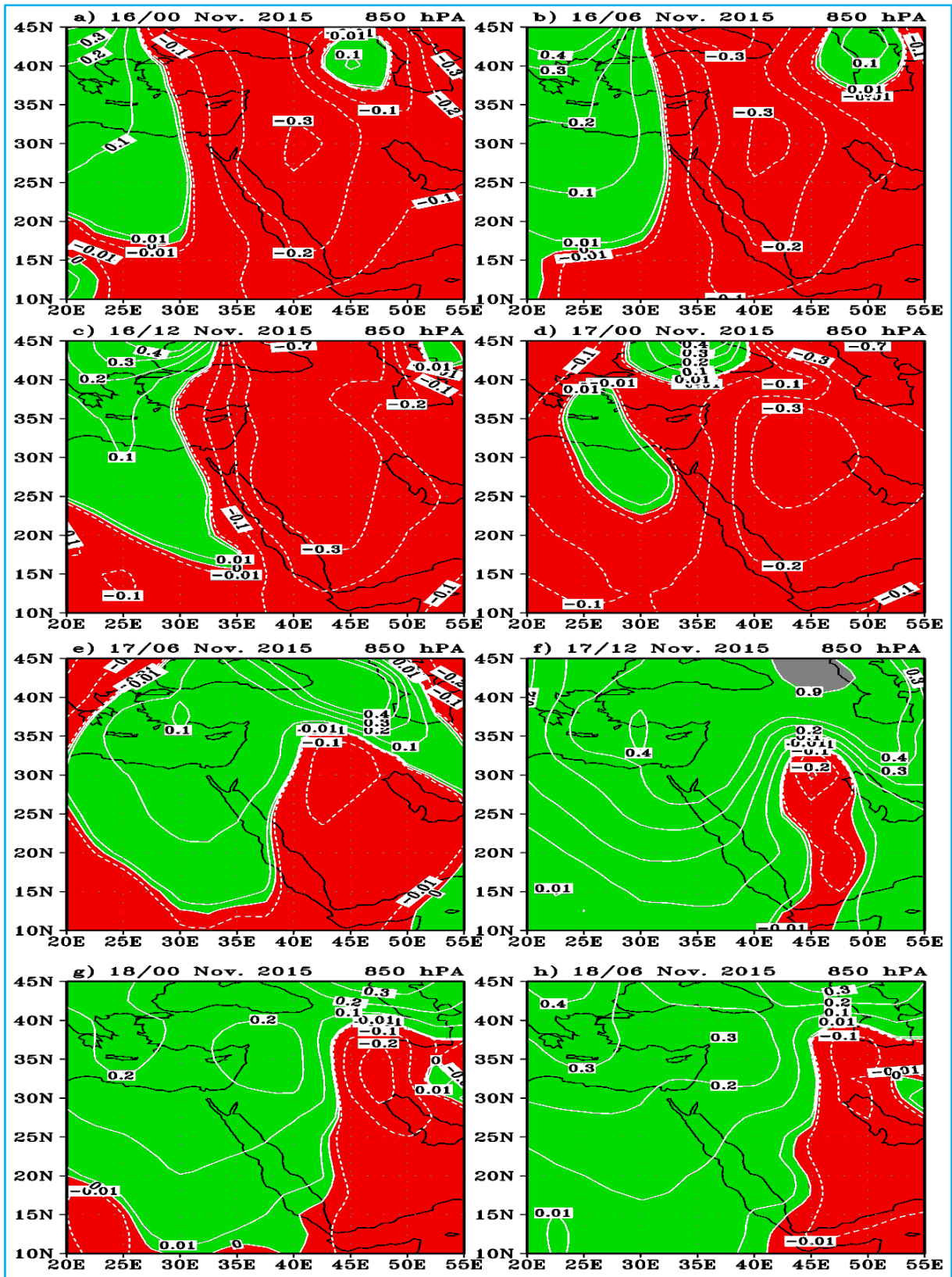
where, the subscript LH refers only to condensational heating. The regions of decreasing static stability are initially unknown and are found through iteration during the ω -equation inversion. The horizontal gradients of heating in this computation were included, which can be considerable in frontal zones. Moreover, the non-conservative generation term is free to feedback on the flow tendency and vertical velocity.

Fig. 11(a-h) displays the EPV generation rates $(DPV/Dt)_{LH}$ at 850 hPa obtained by the above method. It shows that at 16/00 November, there was an area of EPV tendency over east of Mediterranean (northeast of KSA) with a central value of 0.8 PVU/12h, this central value of EPV tendency increased to become more than 1 PVU/12h at 16/06 November. The EPV tendency cell shifted southeast ward to cover northeast of KSA with a central value of 0.8 PVU/12h at 16/12 November. In the following 6 hours (17/00), the EPV tendency cell occupied a larger area and extended southward to reach the south of Saudi Arabia.

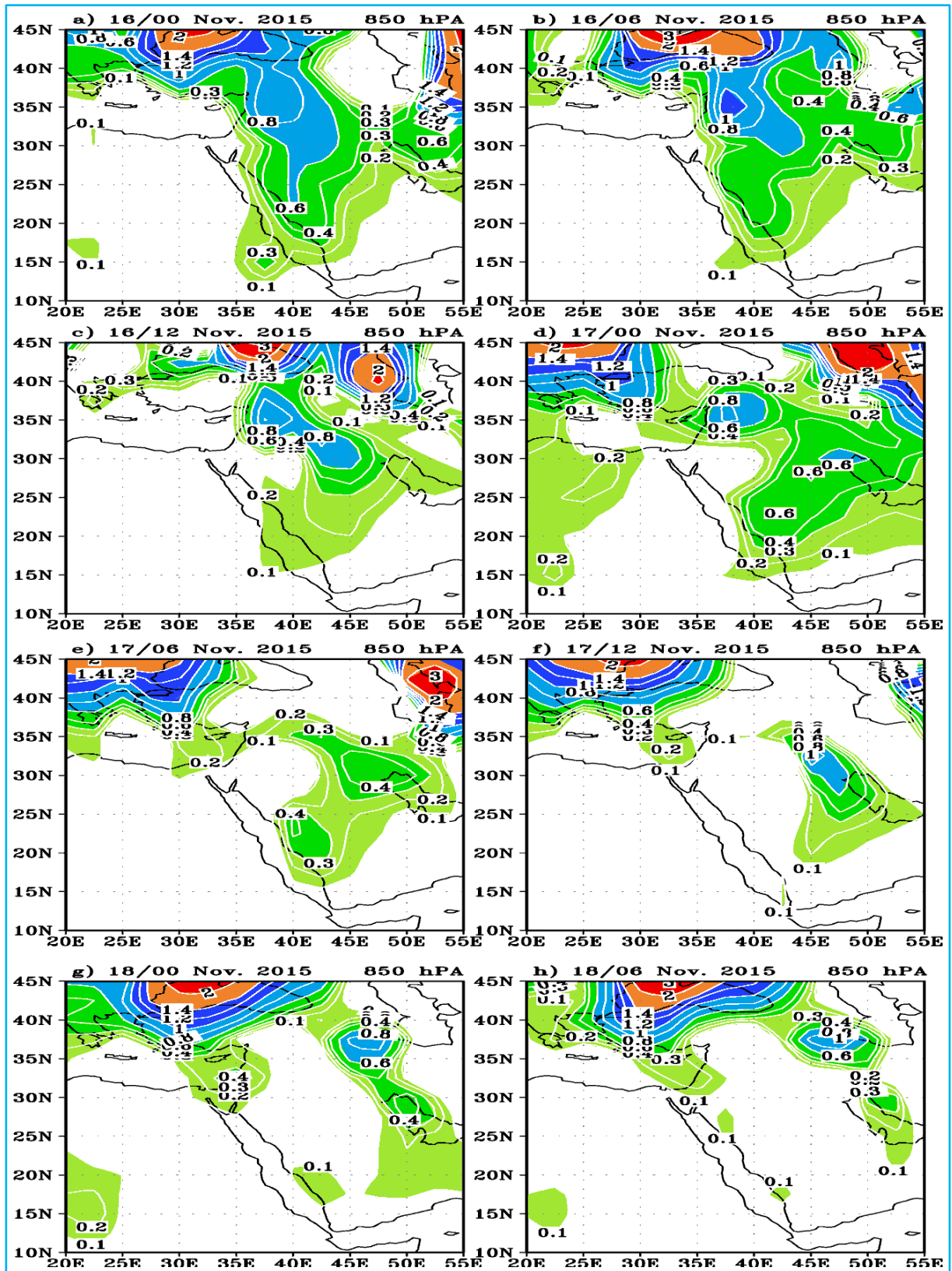
On 17/06 November, the cell of the EPV tendency became closer over the middle of the west of Saudi Arabia, its central value was more than 0.4 PVU/12h, at (21° N, 42° E) and at (27.5° N, 47.5° E). At 17/12 November, the cell of the EPV tendency moved north-eastward and strengthened over the northeast of Saudi Arabia. Its center moved to the north of the Arabian Gulf and had a value 0.5 PVU/12h. In the following two-time steps, the cell of the EPV rate stayed over the northeast of the Arabian Gulf (over Iran) in association with the movement of the studied system. The results of the above discussion signify that the heating region coincides with the location of the EPV generation estimates and the low-level PV anomaly and implies that condensation supplies a large enough source to account for the presence of the low-level EPV anomaly. The low-level diabatically produced PV helps in intensifying



Figs. 9(a-h). Ertel's potential vorticity on 700 hPa, contour interval is 0.1 PVU for 16/00-18/06 UTC November, 2015. $1 \text{ PVU} = 10^{-6} \text{ M}^2 \text{ K Kg}^{-1} \text{ S}^{-1}$



Figs. 10(a-h). Vertical motion at 850 hPa, contour interval is 0.1 Pa S^{-1} , (green color) denote positive values (red color) denote negative values for 16/00 - 18/06 UTC November, 2015



Figs. 11(a-h). Ertel's potential vorticity generation at 850 hPa, contour interval is 0.2 PVU per 6 hr, for 16/00 - 18/06 UTC November, 2009.
Note that only positive generation rates are contoured

the surface thermal wave early in the development of the studied system and in the upper-level wave during the later growth stages.

7. Summary and conclusions

This paper aimed to study the diabatic heating during a case of interaction between midlatitude and extratropical cyclones over east Mediterranean. The relationship between the diabatic processes and the generation of the low-level potential vorticity anomaly during the period of study is also examined. Our case of study was developed over Saudi Arabia on 16, 17 and 18 November, 2015 where north and middle regions of Saudi Arabia received heavy rainfall. Careful examination of the synoptic analysis of our case study illustrate that the mechanism of occurrence of this case begin with the formation of a midlatitude upper level trough that intrudes into northeast of Africa (north Libya). The trough was developed which make the upper and lower tropospheric circulations couple and interact. The existing RST oscillate northward and merge with the midlatitude upper level trough, the ridge over the Saudi Arabia intensifies the persistent anticyclonic flow around the Saudi Arabia at low to midlevel's, together forcing pronounced southerly winds over the Red Sea region. Consequently, the RST extends northward and the associated tropical moist air masses progress over the Red Sea and intrude into Saudi Arabia regions. The enhanced tropospheric instability and dynamical forcing cause ascending motions, promoting the development of convective storms and heavy precipitation. Subsequently, the midlatitude upper level trough dissipates, the RST back southward to its climatological position and the Arabian Anticyclone weakens and the moisture transport from the Arabian and Red Seas is again directed toward equatorial Africa. Calculations of the diabatic heating have been made using the thermodynamic equation in isobaric coordinates. The diabatic heating is strongly associated with the horizontal advected warm air into the region of calculation during the period of study. Also, it is found that the contribution of adiabatic term and the vertical temperature advection are opposite to each other, which produce a strong negative correlation between the patterns of the adiabatic and the vertical temperature advection terms. The contributions of local temperature changes term to the diabatic heating rates are very small comparing with all other terms. The existence of the Red Sea and its neighboring mountains supports the implication that the diabatic heating in the lower atmosphere is mainly due to the latent heat release on the rainy days.

The dynamics of the rainstorm cyclonic system have been also investigated in terms of isobaric potential vorticity. Generally, this approach appears to recognize

the same features for rainstorm initiation: an isobaric PV anomaly at the lower pressure levels with a low-level baroclinic zone. The PV analysis indicated the possible effects at low levels in the central Red Sea (where condensation of water vapor produced a positive lower PV anomaly), an area where the diabatic processes appear to play an important role in convection development. It is also concluded that the location of the low-level PV anomaly and the EPV generation estimates coincides with the heating region, which implies that condensation supports a large enough source to explain the existence of the low-level EPV anomaly. Also, the condensation caused a low-level EPV anomaly that adds directly to the low-level PV anomaly, which in turn is strongly influenced by moisture processes. The low-level diabatically produced PV helped in strengthens of the surface thermal wave early in the convection development and also in the upper-level wave during the later stages of the system's growth.

The behavior of moisture-flux components (divergent and rotational) along with precipitable water content for different tropospheric layers are also examined. The rotational component of the moisture transport brings moisture from two regions; the first which is considered the main region is the Indian Ocean, Arabian Sea, Gulf of Aden and north east of Sudan. The second source region is the Atlantic and Mediterranean Sea.

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