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A case of Khamsin type weather in north Africa

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सार — अधिकांश मामलों में उत्तरी अफ्रीका पर मरुस्थलीय अवदाब एटलस पर्वतों की प्रतिपवनों में बनते हैं ऐसे अवदाब तब पाए जाते हैं जब अटलान्टिक सागर के ऊपर से एक उत्तरी अथवा उत्तर-पश्चिमी पवन धारा एटलस पर्वतमाला की ओर बढ़ती है या जब पश्चिमी भूमध्य सागर से उत्तरपूर्वी पवन इन पर्वतमालाओं की ओर बहती है। जैसा कि चित्र संख्या एक में दिखाया गया है, ये अवदाब अपने पूर्वी स्थानान्तरण के समय बहुत से रास्ते अपना सकते हैं। इन अवदाबों के कारण आमतौर पर भयंकर गर्मी पड़ती है या रेतीले तूफान आते हैं (एलफान्डी 1940, सोलमन 1958)। बसंत में खामसिन मौसम की परिघटना उत्तरी अफ्रीका के क्षेत्र में मौसम के विश्लेषण तथा उसके पूर्वानुमान से जुड़ी मुख्य समस्याओं में से एक हैं। इन अवदाबों के स्रोतों और उष्मा तथा ऊर्जा की पूर्ति के दृष्टिकोणों से हाल के वर्षों में मरुस्थलीय अवदाबों की इन किस्मों के कई संरूपों पर चर्चा हुई है।

यह लेख इन अवदाबों के प्रपथ पर चक्रवात जनन संबंधी गतिविधियों के प्रभाव को पहचानने का एक प्रयास है।

ABSTRACT. In most cases, desert depressions over north Africa form in the lee of the Atlas mountains. Such occurrences are found when a north or northwest air stream from over the Atlantic moves toward the Atlas range, or when a northeasterly wind blows over the western Mediterranean towards these mountains. As shown in Fig. 1, these depressions may follow numerous tracks during their eastward movement. These depressions usually produce severe heat waves and sandstorms (EL Fandi 1940; Soliman 1958). The phenomenon of Khamsin weather in spring is one of the main problems associated with weather analysis and forecasting in the area of north Africa. In recent years, several formulations of these types of desert depressions have been discussed from the point of view of their sources and supply of heat and energy.

The present paper is an attempt to identify the effect of cyclogenetic activity on the trajectory of these depressions.

1. Introduction

The process of cyclogenesis is regarded as the manifestation of the amplification of a small perturbation which is superimposed on an unstable zonal current. For a perturbation to amplify, it is clear that the basic flow must give up potential and/or kinetic energy to the perturbation. In baroclinic instability it turns out that the potential energy of the basic stateflow is converted to potential and kinetic energy of the perturbation. The relationship between the development depressions and the horizontal gradient of temperature (horizontal baroclinicity) has been discussed by numerous meteorologists from various viewpoints.

2. The case under study

The phenomenon of Khamsin weather in spring is one of the main problems in weather forecasting in the area of north Africa. Khamsin is a southerly wind blowing over Egypt in front of depressions passing eastwards along the Mediterranean or north Africa, while pressure is high to the east of the *Nile*. As this wind blows from the interior of the continent, it is hot and dry and often carries much dust. Khamsin is frequent from April to June. These desert depressions usually formed as a lee wave of the Atlas mountains either when north or northwest air streams from the Atlantic move towards the Atlas range, or when northeasterly winds blow over the western Mediterranean towards these mountains. There are different patterns for the classification of Khamsin behaviour given by Kung and Baker (1975). In this study, we are only considering type (C_1) according to that classification (a closed low at 500 mb and developing cyclonic circulation/low pressure area at the surface).

In view of the scarcity of observations in this region, the study of desert depressions is not easy. A particular case of Khamsin weather, 10-14 March 1976, was selected for the purpose of this study when the surface and upper air data were most complete.

3. Formulation of the problem

On the basis of Lytfilli (1966), the thermodynamic equation can be written as :

$$\frac{dT}{dt} - \frac{\gamma_a}{gp}\frac{dp}{dt} = \frac{E}{\rho c_p} \tag{1}$$

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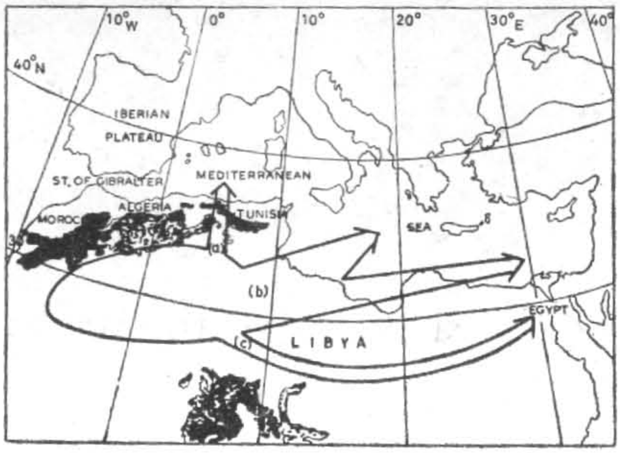

Fig. 1. Tracks followed by depressions originated south

In local terms of temperature and pressure, the equation becomes :

$$\frac{\partial T}{\partial t} - \frac{\gamma_a}{gp}\frac{\partial p}{\partial t} + (\gamma_a - \gamma)\, w - A_T = \frac{E}{\rho\, c_p}$$

$$\Omega_g = \frac{\partial}{\partial x} v_g - \frac{\partial}{\partial y} u_g \simeq \frac{1}{fp} \triangle p$$

$$\frac{\partial \triangle T}{\partial t} - \frac{\gamma_a}{gp}\frac{\partial \triangle p}{\partial t} + \triangle\, (\gamma_a - \gamma)\, w - \triangle A_T = \frac{\triangle E}{\rho c_p}$$

Put $r = \gamma_a - \gamma$

$$\frac{\partial \triangle T}{\partial t} - \frac{f\gamma_a}{g}\frac{\partial}{\partial t}\,\Omega_g - \frac{f\gamma_a}{gp}\,\Omega_g\frac{\partial p}{\partial t} -$$

$$- \triangle(A_T - rw) = \frac{\triangle E}{\rho\, c_p}$$

The term due to local variation is neglected by order of magnitude and assuming $\triangle E = 0$ for simplicity :

$$\frac{\partial \triangle T}{\partial t} - \frac{f\gamma_a}{g}\frac{\partial \Omega_g}{\partial t} - \triangle(A_T - rw) = 0 \tag{2}$$

From definition of thermal wind

$$\frac{\partial v_g}{\partial z} = \frac{g}{fT}\frac{\partial T}{\partial x}$$

$$\frac{\partial u_g}{\partial z} = \frac{g}{fT}\frac{\partial T}{\partial y}$$

By differentiation and addition we get :

$$\triangle T = \frac{fT}{g}\frac{\partial \Omega_g}{\partial z} \tag{3}$$

Then from Eqns. (2) and (3) we get :

$$\frac{\partial}{\partial z}\frac{\partial \Omega_g}{\partial t} = \frac{R}{fH_0}\triangle\,(A_T - rw) \tag{4}$$

where

Ω_g is the geostrophic vorticity.

R is the universal gas constant.

$f = 2\,\omega \sin\,\theta$ is the coriolis parameter.

w is the vertical velocity.

H_0 is the height of the pressure level (RT/g).

Temperature advection (A_T) can be expressed as :

$$A_T = \frac{g}{T} J(H, T) - \frac{q}{c_p} \tag{5}$$

A_T is the temperature advection.

g is the acceleration of gravity.

T is the temperature at the pressure level.

H is the geopotential height.

q is the non-adiabatic effects.

$J\,(H,T)$ = Jacobian of H and T.

$r = \gamma_a - \gamma$ is a parameter due to stability of the atmosphere.

γ_a is the dry adiabatic lapse rate.

γ is the actual lapse rate.

$\triangle$ represents Laplacian

ζ = Relative vorticity.

From Eqn. (4) the dynamical cyclogenesis-anticyclogenesis can be defined by the aid of the advection of temperature, vertical velocity of air and atmospheric stability. These in turn are directly related to the formation of the atmospheric fronts.

It is obvious that the variation of (A_T, w, r) in the vertical are considerable, while the horizontal variations

TABLE 1

Cyclogenesis factor (units 10^{-12}/sec^2)

Date (March 1976)	Time (GMT)	C_{sfc}	C_{850}	C_{700}	C_{500}
10	0001	+76.618	+5.990	−0.475	−4.938
	1200	+99.288	+3.429	+1.842	+0.605
	0001	+46.055	+4.625		
11	0001	−50.507	−4.016	−0.910	−.0.725
	1200	−23.590	−1.589		
	1200	+9.383	−11.977	−4.751	−3.981
12	0001	−6.159	−8.016	−5.565	−3.062
	1200	+4.399	+1.506	−4.950	−5.512
13	0001	+1.803	+7.988	−4.926	−6.064
	1200	−6.331	−2.213	−4.190	−0.498
14	0001	−25.151	−4.673	−5.865	−3.324
	1200	−76.853	−18.249	−19.846	−12.094

are so small compared with that in the vertical. Therefore, Eqn. (4) shows the local variation of relative vorticity with height, what we call frontal cyclo-anticyclogenesis.

If we consider cyclogenetic processes in the troposphere, Eqn. (4) can be written as :

$$\frac{\partial \zeta \bar{z}}{\partial t} - \frac{\partial \zeta z}{\partial t} = \frac{R}{f_0 H_0} \int_{z}^{\bar{z}} (A_T - \triangle rw)\, dz \qquad (6)$$

where $\frac{\partial \Omega z}{\partial z} = \frac{\Omega \bar{z} - \Omega z}{\triangle z}$ and $\bar{z}$ = height of tropopause.

As a boundary condition, if one disregards the variation of $\zeta_{\bar{z}}$ with time, then one can derive :

$$\left(\frac{\partial \zeta z}{\partial t}\right)_0 = -\frac{R}{f_0 H_0} \int_{z}^{\bar{z}} (A_T - \triangle rw)\, dz \qquad (7)$$

From this equation, the rate of change of vorticity at any level can be computed.

From scale analysis, the temperature advection is in the order of 10^{-12} while the term (rw) is of the order of 10^{-13}. From Eqn. (7) we can see that the second term on the R.H.S. is small as compared with the first one, which means that A_T can give a good approximation for the cyclogenetic factor (C.F.).

In the present study, A_T was calculated by the aid of Eqn. (5) for the sake of simplicity, neglecting the second term due to the effect of the non-adiabatic heating. The computations were performed at the centre of the surface depression during its life time, and also at the centre or the lows at the isobaric levels 850, 700 and 500 mb during the same period.

4. Discussion

The results are summarized in Table 1. From these values it can be interpreted that :

(a) Cyclogenetic factor (C.F.) has a large positive value in the lower layer (surface to 850 mb level) during the passage of the depression over the north Africa coast.

When the surface depression crossed the coast, C.F. began to exhibit negative values in the lower layer. At the same time the depression suffered a splitting at these two levels.

(b) C. F. began to take positive values again in the lower layer when the depression approached the Cyprus Island and south coast of Europe.

(c) At the upper levels (700, 500 mb), cyclogenesis always exhibited a negative value whenever the lows at these two levels did not split.

(d) After reaching the eastern coast of the Mediterranean and during the travel over land east of the Mediterranean, C. F. is decreased rapidly at all levels.

(e) The movement of the depression in areas of positive cyclogenesis was slowly [from 0001 GMT on the 10th to 0001 GMT on the 11th and from 1200 GMT on the 12th to 0001 GMT on the 13th] (*see* Table 1). While the movement of the depression in areas of negative C.F. is more rapid [from 0001 GMT on the 11th to 1200 GMT on the 12th] (Table 1).

Such findings are in complete agreement with our day to day experience in analysis. When depressions approach the area south of the Gulf of Sidra or the area of Cyprus Island, its motion becomes slower than in the areas before and after.

5. Results

From the previous remarks, we can conclude the following :

(*i*) Cyclogenesis depends largely upon the orographic effect. Because it took positive values while the centre of the depression was approaching the coast line either from land to water (at the north coast of Africa) or the reverse from water to land (south coast of Europe). C. F. became largely pass on crossing the north African coast where the subtropical front was located. This can be interpreted as the result of a great difference in temperature caused by the relatively colder sea water and the hot land surface of the Sahara, which is thought to be the main factor for the deepening of these depressions over that area.

(*ii*) It appears that the main factor of steering for these depressions is their attraction towards the areas of positive C. F. At the same time this is the main factor for retarding the depressions at the two areas [Gulf of Sidra and Cyprus Island].

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