TECHNICAL REPORT No. 25

ON THE ATMOSPHERIC FACTORS AFFECTING THE LEVANTINE SEA

by

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C O N T E N T S

EDITOR'S FOREWORD

This report was written during a visit of the author to ECMWF in September 1980. The report does not directly deal with medium range numerical weather prediction; the problem of air-sea interaction is discussed specifically in relation to the Mediterranean region. The report contributes to our understanding of how these processes may affect the atmosphere on the time scale of one month, a time scale which is relevant to extended range numerical weather prediction.

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1. INTRODUCTION

The Levantine Sea is the eastern part of the Mediterranean located to the east of the island Crete and separated from the Aegean Sea by the Crete-Rhodes island arc. The Levantine Sea is bounded by the coasts of Turkey in the north, Syria, Lebanon and Israel in the east and Egypt and Libya in the south.

The Mediterranean Sea is rather a unique area of the world where the dynamics of the atmospheric and marine motions are largely governed by the strong two-way interaction taking place at the air-sea interface. Furthermore, due to the semi-enclosed geometry of the basin, the topographic features of the surrounding land masses and the sea bottom play significant roles in determining the types of motions of the atmosphere and the sea, respectively. In comparison to the importance of the climatological, ecological and dynamical processes taking place in the region, limited success has been achieved by research in the past in providing a more complete understanding of these processes, although some major breakthroughs have been made on particular phenomenological aspects. The atmospheric and oceanographic research in the Mediterranean is therefore a developing frontier, moving at an accelerated pace during recent years. The eastern Mediterranean in particular is open to such investigations since the available information on it is very limited compared to the western parts of the basin.

Although two-way interactions are important in the region, only the atmospheric inputs to the sea and their large-scale effects in creating water masses will be discussed in Section 2, with special attention given to the eastern Mediterranean.

2. FORMATION OF WATER MASSES IN THE MEDITERRANEAN

The semi-enclosed Mediterranean basin which is connected to the Atlantic ocean through the Strait of Gibraltar can be classified as an arid sea (Neumann and Pierson, 1968) in which the water loss by evaporation into the atmosphere exceeds the water gain by runoff from land and rainfall over the sea. This sea is often called the Asian Mediterranean, since similar situations occur in other parts of the world such as the Carribbean sea (American Mediterranean). This contrasts with humid semi-enclosed seas such as the Black Sea and the Adriatic Sea in which the rainfall plus runoff exceeds evaporation. The excess of evaporation in an arid semi-enclosed sea sets up a thermohaline circulation in which the surface waters entering the sea become denser as they move away from the entrance and the ensuing sinking motions generate a deep thermohaline circulation, with the deep waters leaving the sea through the opening. However, there must be an excess of water entering through the opening to maintain the mass deficit through net evaporation. The net evaporation deficit in the Mediterranean has been calculated as 1929 km^3/year by Lacombe (1977). If there was no net inflow through the Gibraltar Straits, the Mediterranean sea level would be decreased by 0.76 m/year. In a humid, deep semi-enclosed sea the reversed balance of surface mass flux limits the thermohaline circulation to the surface layers above the level of the sill at the opening, with little or no motion at deeper layers, such as in the case of the Black Sea.

The surface waters flowing in an easterly direction from the Gibraltar region generate mean currents as influenced by the variability of mean winds at the surface and the interior currents which are in turn influenced by the thermohaline circulation and bottom topography. (Wüst 1960, Lacombe 1972, Gerges and Karam, 1978). The deeper circulation is more complex due to the thermohaline structure shown in Fig. 1 (Wüst, 1960). Several distinct water masses are formed which can be identified by salinity and temperature values characteristic of the climatological conditions at which they are formed, and these deeper waters return to the Atlantic as a sub-surface current through the Gibraltar Straits. This vertical circulation is in some sense similar to an estuarine circulation with relatively much larger scale (Gerges, 1972). The residence time corresponding to this circulation is estimated as 97 years (Lacombe, 1977).

The thermohaline circulation shows some seasonal variation. In summer, the surface water flowing towards the east is subject to strong solar radiation and high evaporation, and in the eastern basin reaches extreme values of temperature and salinity such as $T = 30^{\circ}C$ and $S = 39^{o}/oo$. But due to the high surface temperatures, the surface waters are less dense than the underlying waters and mixing is suppressed by a strong seasonal thermocline. In winter, the surface waters transformed during the summer season are subject to dry and cold winds associated with cold outbreaks at the northern borders of the Mediterranean and at certain regions sinking motions associated with strong vertical mixing, cause the formation of distinct water masses at intermediate and deep layers.

One of the most dominant water masses affecting the whole Mediterranean is the Levantine Intermediate Water (LIW) named after its region of formation. This water mass, identified by a salinity maximum at intermediate depths, is primarily formed in the winter season and flows westward at depths of 200 - 400 m, as shown in Fig. 1. Due to the presence of this warm and saline tongue of water originating in the Levantine Sea, a three layer stratification exists in the whole Mediterranean, with low salinity Atlantic type water near the surface, warmer but more saline waters at intermediate depths, and cold and saline waters below this layer. While the LIW flows in a westerly direction, the salinity at its core region is diluted by mixing, but the flow can be identified as far from its source region as the Gibraltar undercurrent. The flow is largely affected by the bottom topography and the various sills along its course and therefore branches out into the smaller basins of the Mediterranean (Wüst, 1960). The Mediterranean basin is divided into two major basins by the Sicilian Sill. The deep waters of the western basin are formed along the region to the south of France through mixing of the surface waters with the LIW, whereas the deep waters of the eastern basin are formed as a result of the mixing of the Adriatic outflow with the LIW (Wüst, 1960). Therefor the LIW plays a crucial role in the physical and ecological characteristics

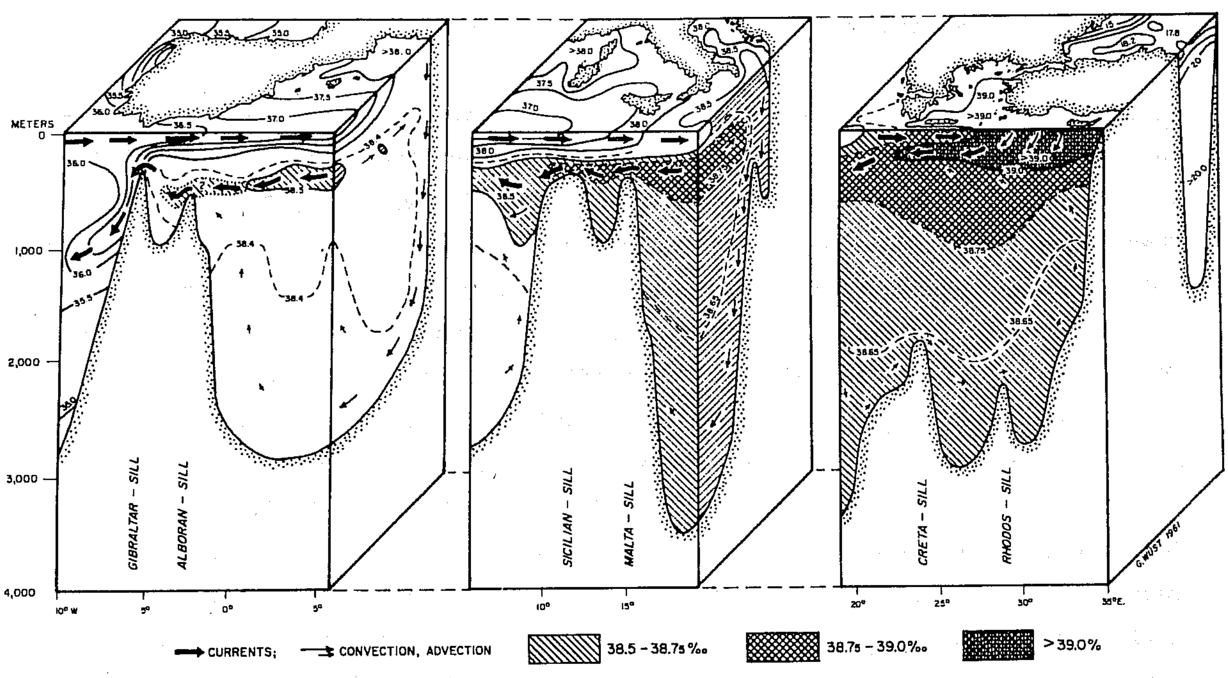


Fig. 1 The vertical circulation in the Mediterranean Sea (after Wüst, 1961).

of the Mediterranean. For example, the transport of nutrients by the sinking motions and by the Levantine intermediate current towards the west is responsible for the poor fisheries activity and primary production observed in the Mediterranean, and especially in the Levantine Sea (McGill, 1960).

The striking coincidence of areas of water mass formation and the regions affected by outbreaks of cold and dry air masses reaching the Mediterranean from the northern continental regions has been noted by Wüst (1960) and Lacombe and Tchernia (1974). As will be shown in the next Section, the unique topography of the northern coasts of the Mediterranean has a crucial influence on the locations of such centers of water mass formation. Furthermore, there is also a correspondence between these locations and the three well known regions of maximum cyclogenesis in the Mediterranean, probably not because of the direct influence of the surface winds associated with cyclones, but due to the interaction of the synoptic scale disturbances with meso-scale topographic channelling effects. The incursions into the Mediterranean of cold and dry polar continental air masses through the numerous valleys on the northern borders of the sea are especially forceful in the neighbourhood of cyclogenetic regions, and the enhanced evaporational instabilities and mixing in the surface waters give rise to water mass formation.

In the western Mediterranean there is a close relationship between the occurrence of Mistral and Tremontana winds blowing through the valleys located between the Pyrenees and Alps and the Genoa cyclogenesis (Crowe, 1971, Reiter, 1975). These winds in turn create excessive evaporation and mixing in the surface waters which become subject to different phases of transformation in the warm and cold seasons and finally lead to the formation of deep waters of the Western basin (Gascard, 1978). During the early preconditioning phase, warm and saline waters are formed at certain spots around which a cyclonic flow pattern is developed in the surface layers of the sea. In the following baroclinic instability phase, water masses with different rates of stratification are found adjacent to each other, separated by many frontal features. As a result of the baroclinic instability mechanism, descending and ascending motions associated with slanted convection take place and more homogeneous water masses are developed at the centers of the circulation cells. This is followed by the violent mixing phase in which rapid mixing triggered by Mistral winds occurs in a few days over core regions of a few kilometers in size and the mixing reaches down to depths of more than 2000 m. (Stommel, 1972, Gascard, 1978).

Along the Dalmatian coast to the north of the Adriatic Sea, the Bora winds in winter are triggered either by western and central Mediterranean atmospheric cyclones or by the eastern flanks of the advancing surface fronts that are intercepted by the Alps (Crowe, 1971). Bora-induced mixing in the shallow Adriatic basin reduces the temperature of the Adriatic outflow (produced by the excess of rainfall to evaporation in this basin) which in turn forms the deep waters of the eastern Mediterranean basin by mixing with the Levantine Intermediate Water at the Ionian Basin (Wüst, 1960, Zore-Armanda, 1972, McGill, 1960).

The LIW itself is formed along the Southern coasts of Asia Minor where cold and dry outbreaks surging into the Mediterranean are effective during winter months. However, there appears to be more than one source region responsible for the formation of LIW. Two source regions are detected on both sides of Rhodes, other source regions are located to the north of Cyprus, and a source region of secondary importance appears near the Egyptian coast (Wüst, 1960, Morcos, 1972, Özturgut, 1976). Only after reaching the Ionian basin does the LIW become more uniform (Morcos, 1972). However, it has been shown by Özsoy et al (1980) that probably the most important source region is located along the Turkish coast in the vicinity of the Göksu river valley and secondly near the Bay of Antalya where channelling effects of topography are important in generating forceful winds in the winter season. Especially near the Göksu valley a local northerly wind system called Poyraz is triggered either by cyclones generated near the Cyprus cyclogenesis area or by passing cyclones (Ataktürk, 1980). Time series of wind stress, relative humidity and sea water temperature are displayed in Fig. 12 for a station within the region of influence of Poyraz winds. The water temperature within the surface layer indicates stepwise drops of rather large magnitude (1 - 2°C) during each Poyraz event. In fact, these observed stepwise drops in water temperature are largely responsible for the seasonal variation of sea water temperature within a total yearly range of about 12°C. In winter the sea water is still warmer than the air by about 10°C in the region. The strong cold and dry winds blowing over warm sea water during Poyraz cause the evolution of the surface waters towards higher densities during winter months through intensified mass and latent heat fluxes associated with evaporation, and the sinking of the dense waters ensues following the month of February (Özsoy, et al (1980).

The above observations lead to a similarity between the three regions of the Mediterranean where atmospheric inputs significantly affect the marine environment, although one must be cautious not to extend the analogy too far to the exact meso-scale mechanisms acting at specific geographical regions with different types of climatic regimes. Therefore, a review of the general behaviour of the atmosphere in the Mediterranean region, with some discussions of regional characteristics seems highly necessary. This is the purpose of the next Section.

3. ATMOSPHERIC CHARACTERISTICS INFLUENCING THE MEDITERRANEAN SEA

Weather prediction in the Mediterranean is a challenging problem for forecasters in the area and often requires a good knowledge of the intricate mechanisms responsible for the general behaviour of the atmosphere in this region (Reiter, 1975, Karein, 1979). As pointed out earlier, the Mediterranean meteorology has rather unique properties, as compared to the other areas of the world, due to the air-sea interactions and the surrounding topography.

3.1 Planetary and synoptic scale phenomena

Due to its semi-enclosed geometry, the Mediterranean waters are much warmer than the atmosphere during winter and the maximum yearly rainfall is observed, in the Mediterranean region, in the winter season, associated with frequent occurrence of cyclonic perturbation systems. The baroclinicity present between the warm sea region and the northern continental mass occupied by cold air masses of polar origin is thus enhanced in winter months. As a result of the confluence of winds from Europe and Africa into the Mediterranean low pressure trough, the average position of the polar front reaches the Mediterranean and forms what is often called the Mediterranean front (Fig. 2).

The Mediterranean frontal zone is third in importance and persistency in the Atlantic area preceded by the Atlantic Polar and Atlantic Arctic fronts (Peterssen, 1956). In this environment frequent cyclogenesis occurs in the region. Of the 76 cyclones observed annually, only 7 enter the Mediterranean from outside, with 69 being formed within the basin (Karein, 1979, Mediterranean Pilot, 1976). The majority of these cyclones are formed south of the Alpine mountain chain in the Genoa cyclogenesis region, followed next by the region to the south of the Atlas mountains. Only a few are formed in the central and eastern Mediterranean region, and many of the frequent cyclones observed in the Ionian and Levantine basins enter these areas from the western cyclogenesis regions (Fig. 3).

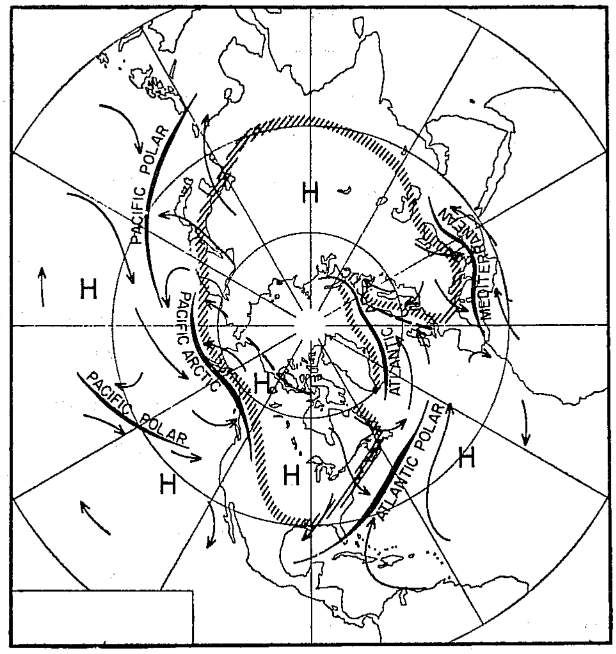


Fig. 2 The principal frontal zones in winter. H = high-pressure areas. Hatchings indicate zones of maximum temperature gradient. Arrows are skeleton indications of wind. From 'Weather analysis and Forecasting' by Peterssen. Copyright (c) 1956 by the McGraw-Hill Book Company Inc. Used with the permission of McGraw-Hill Book Company.

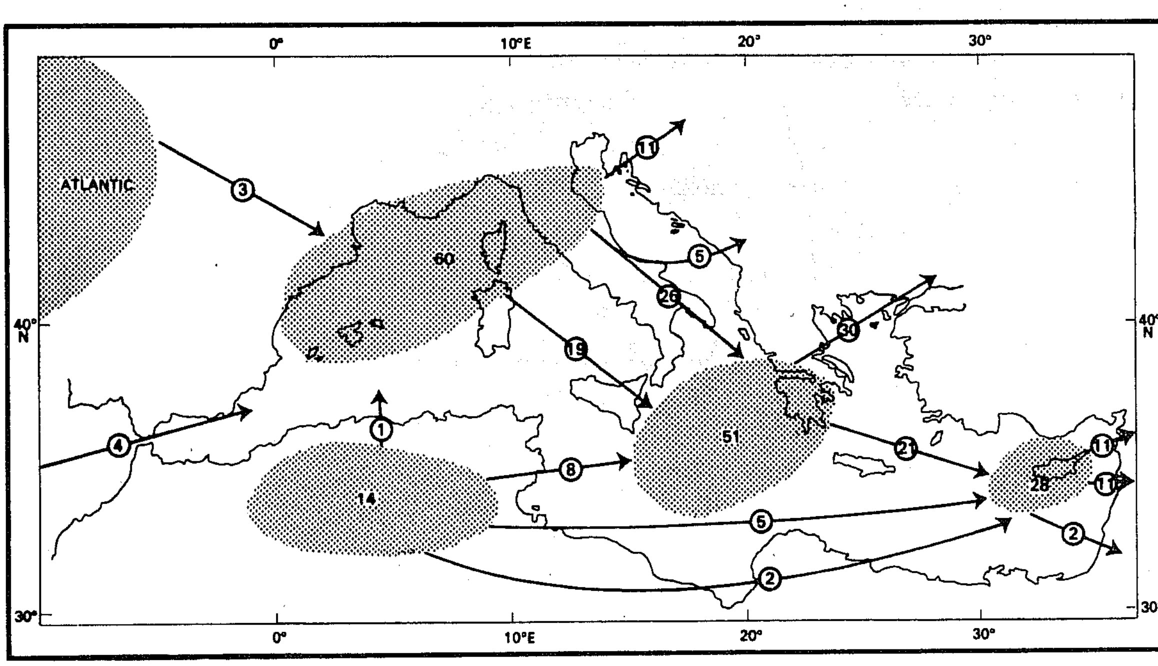


Fig. 3 Tracks of Mediterranean depressions. Numbers indicate average annual frequencies. (After Mediterranean Pilot, 1976). Produced from portion(s) of Mediterranean Pilot Vol V 6th Edition (1976) with the sanction of the Controller, HM Stationary Office and of the Hydrographer of the Navy.

The cyclones on the semi-permanent Mediterranean front travel eastward, with some of them branching to the northeast into the Black Sea region and arriving at southwestern Siberia, and others following the southern coast of Asia Minor and arriving in India (Peterssen, 1956). In fact, whenever a cyclone appears over the central Mediterranean and the Aegean region, it is a major problem for forecasters to predict which of the branches it will follow (see Fig. 3).

The interactions of the polar front jet (PFJ) and the subtropical jet (STJ) are also very important in the Mediterranean. Reiter (1975) points out that the mean position of the STJ in winter contributes to a three-wave pattern surrounding the globe (Fig. 4). Peak winds and anticyclonic curvature of the STJ are observed at three regions: to the east of the Rocky Mountains, to the northeastern part of Africa near the Mediterranean and to the east of the Himalayas. The Rocky Mountains and the Himalayas are responsible for the wave pattern of the STJ, whereas a third maximum occurs near the Mediterranean in resonance with these two waves, since the mountains in this region do not form a major obstacle for the STJ. Upper winds due to the Mediterranean maximum

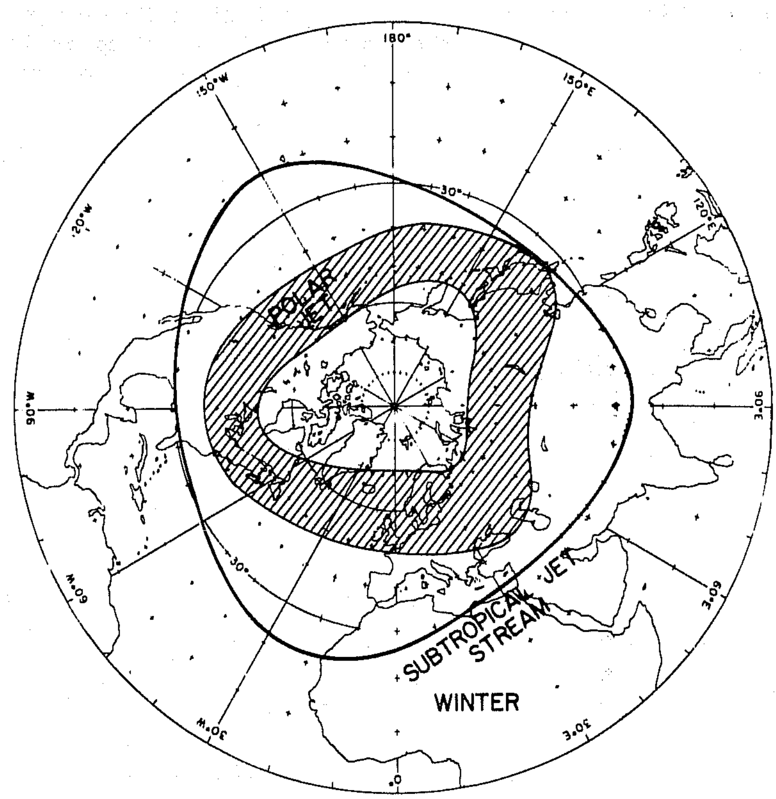


Fig. 4 Mean axis of subtropical jet stream during winter and area (shaded) of principal activity of polar-front jet stream. (After Palmén and Newton, 1969. Reprinted with the permission of Academic Press, Inc. Figure originally taken from H. Riehl, 1962; jet streams of the atmosphere. Tech. Rep. No.32, Dept.Atmospheric Sci., Colorado State Univ., Fort Collins, Colorado, USA, pp 117.)

is not as pronounced as the others (Reiter, 1975). In contrast with the STJ, the mean position of the PFJ has an opposite sense of curvature. At the above mentioned locations, the PFJ forms a trough, while the STJ forms a ridge shape, hence the STJ-PFJ interaction (Fig. 4). Reiter and Nania (1964) have demonstrated that the interaction of the STJ with the PFJ over the United States during periods of meridional flow gives rise to strong cases of cyclogenesis to the east of the interaction area. Similar interaction phenomena are known to exist near Japan. However, Reiter (1975) comments that although no such cases have been demonstrated in the Mediterranean it is expected that such interactions should often occur in the region. In fact, cyclogenesis as a direct result of the interactions of the two jet streams has since been shown by Karein (1979) to take place in the central and eastern Mediterranean regions.

The STJ is not a steady feature as implied by Fig. 4, but it is relatively more steady as compared to the more or less erratic behaviour of the PFJ. There are large seasonal variations of the STJ but during a season it has relatively constant strength and position with only small day-to-day variations, due to the relative constancy of the Hadley circulation in low latitudes (Palmén and Newton, 1969).

The PFJ however, is highly migratory. This is shown in Fig. 5 where the mean and eddy kinetic energy distributions are shown in the winter and summer months. The maximum mean energy is concentrated over U.S. and Japan near the regions of PFJ-STJ interaction. However, the maximum over north-east Africa is weaker and situated more to the south, which better corresponds to the average position of STJ, the reason being the larger latitude range and higher variability in the migration of PFJ over Europe. The eddy kinetic energy thus shows large values over Europe and the Mediterranean regions. In summer the circum-hemispheric system breaks down due to the summer migration of the Intertropical Convergence Zone to further north (Reiter, 1975).

The interactions of the PFJ and the STJ are often associated with the formation of blocking highs and the resulting strong meridional flows. Blocking highs are most commonly observed over Europe, and to the westerly flow they appear as very effective barriers. According to the typical weather classifications compiled by the German Research Institute for Extended Weather Forecasting in 1941 - 1943 (as a forecast tool before computer predictions), more than half of the typical weather patterns occurring in Europe are characterized by blocking highs. They occur most commonly between January and May and can be classified as stationary, retrogressive or progressive. After the formation of a blocking ridge, a high pressure

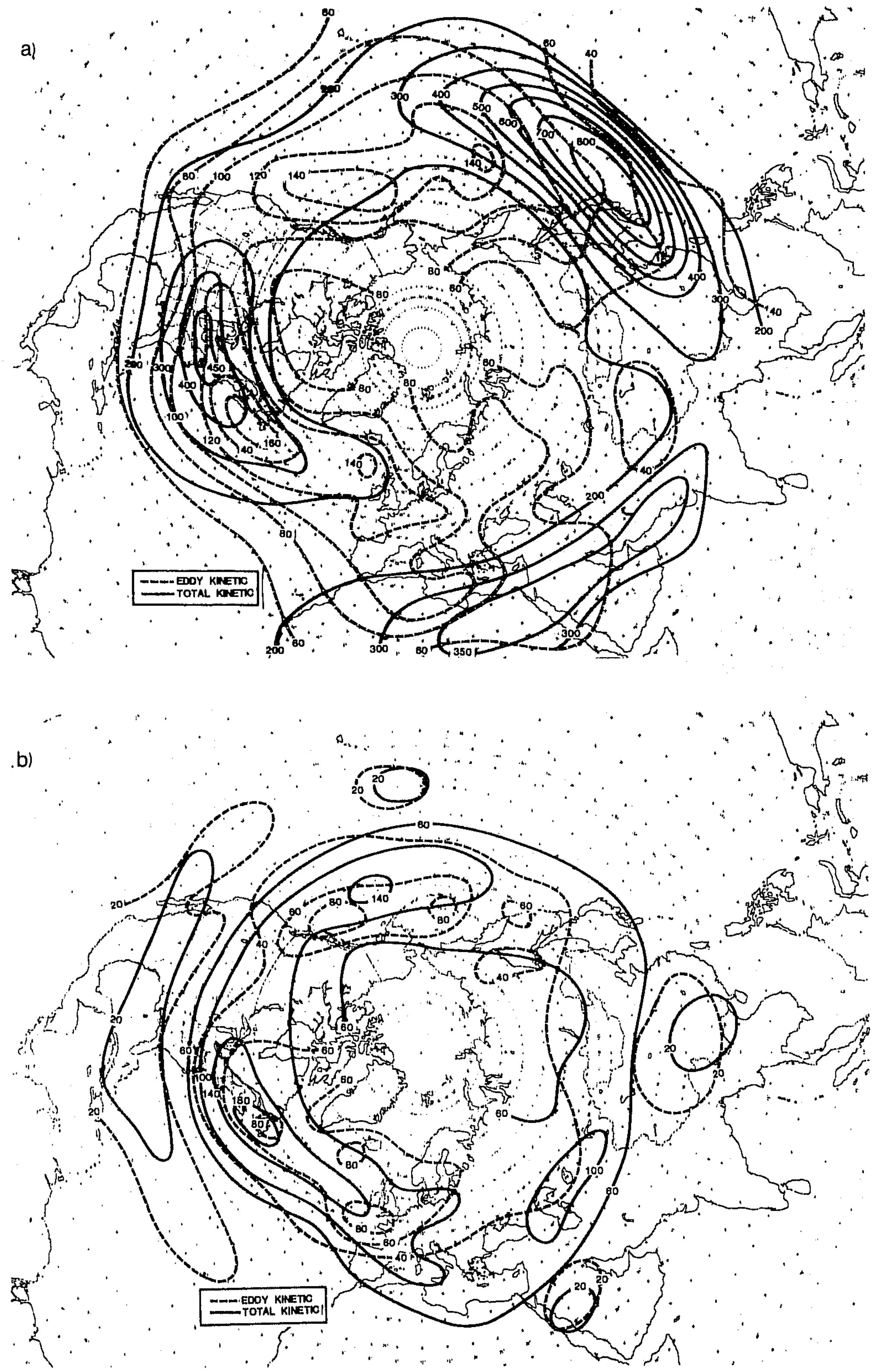


Fig. 5 Distribution of mean total (solid lines) and eddy kinetic energies (dashed lines), (kg m^{-1} sec^{-2}) at 300 mb, a) January, b) July (After Reiter, 1975).

vortex becomes separated and forms a so-called "omega-block". The mean persistence period of these blocks is around 14 days, with maxima of about 34 days (Reiter, 1975). At certain times, blocking highs can persist for a whole season with small interruptions, resulting in extreme drought conditions.

Orographic influences are most significant within the framework of the large-scale characteristics described above. The role of large-scale mountain chains such as the Rockies and the Himalayas have been already mentioned. Various theories are offered to explain the development of frequent blocking highs over Europe, but it is also possible that these features occur as a resonant response to the presence of large-scale orography to the west of STJ maxima in the two regions mentioned above (Rockies and Himalayas).

Another large-scale orographic feature of importance is the presence of the east-west oriented Eurasian mountain chain extending from the Pyrenees to the Himalayas, which, by preventing the free north-south advection of lower-tropospheric air, introduces a large meridional contrast in thermal and moisture fields. Mintz (1965) has found that the occurrence of the Siberian High and the assymetry of the northern hemispheric circulation are mainly caused by this mountain chain (after Palmén and Newton, 1969).

The topography of the continental masses surrounding the Mediterranean is shown in Fig. 6. The Pyrenees, Alps, Atlas Mountains, Caucasian Mountains, (together with their extension into eastern parts of Asia Minor), and the Taurus Mountains form the major obstacles with elevations exceeding 2500 - 3000 m. Besides these elevations, the Iberian Peninsula, the western Sahara, the Balkan Peninsula and Asis Minor cover large areas with elevations above 1000 m. The major mountain chain along the northern borders of the Mediterranean is crossed by numerous gaps in the form of river valleys or mountain passes (Fig. 7). These gaps play an important role in the interaction of synoptic scale motions with the meso-scale motions, as will be explained later.

Reiter (1975) has suggested that the blocking high commonly observed near the Iberian Peninsula may induce the southerly passage of the PFJ over the Alps and may subsequently influence the Genoa cyclogenesis through an interaction with the STJ. More recent research has shown the opposite case to be possibly closer to the truth. Genoa cyclogenesis is formed due to the sharpened baroclinicity over the Alps by the retardation of the north-easterly low-level flow. In a

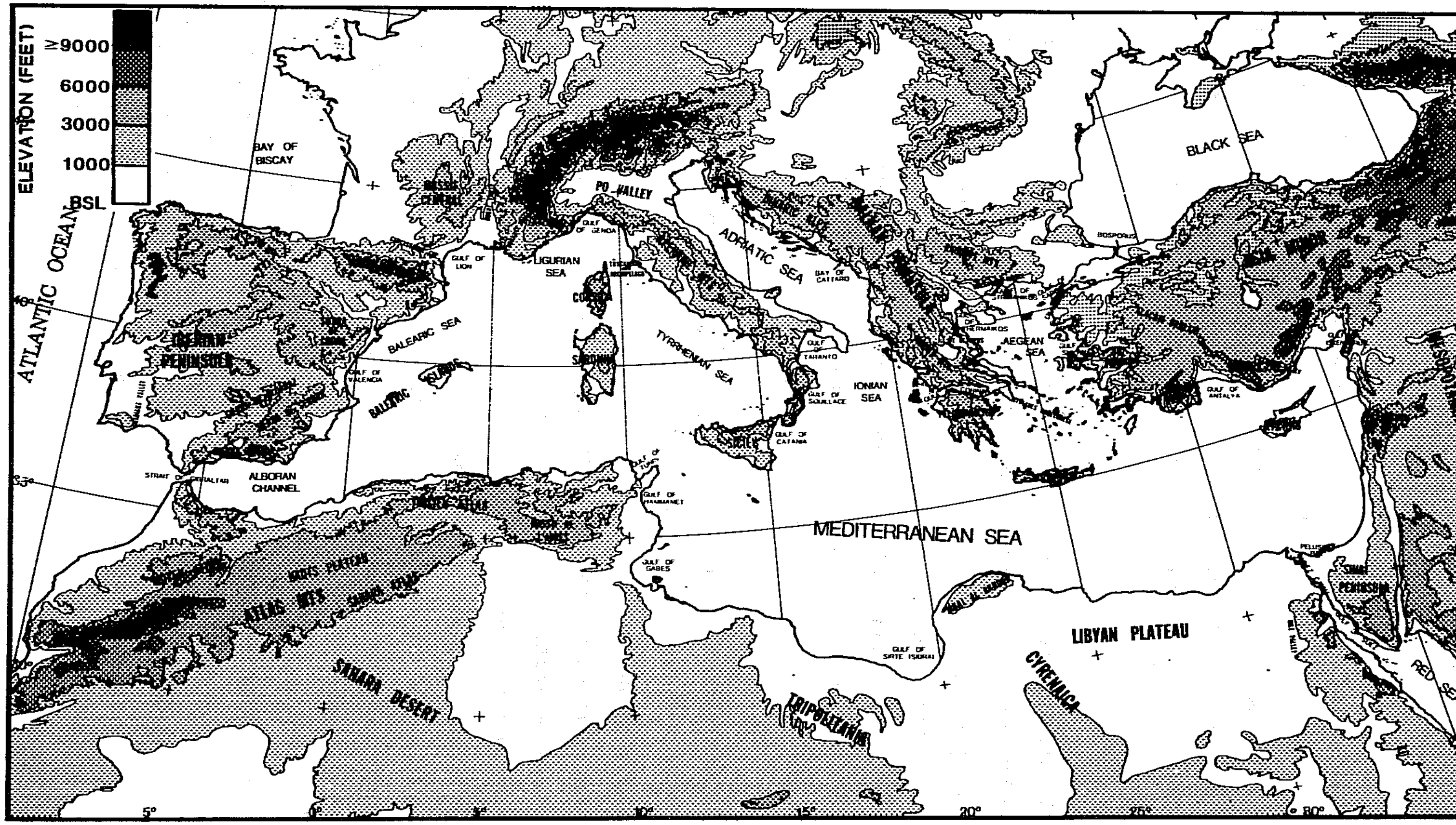


Fig. 6 Topography surrounding the Mediterranean Sea. (After Reiter, 1975)

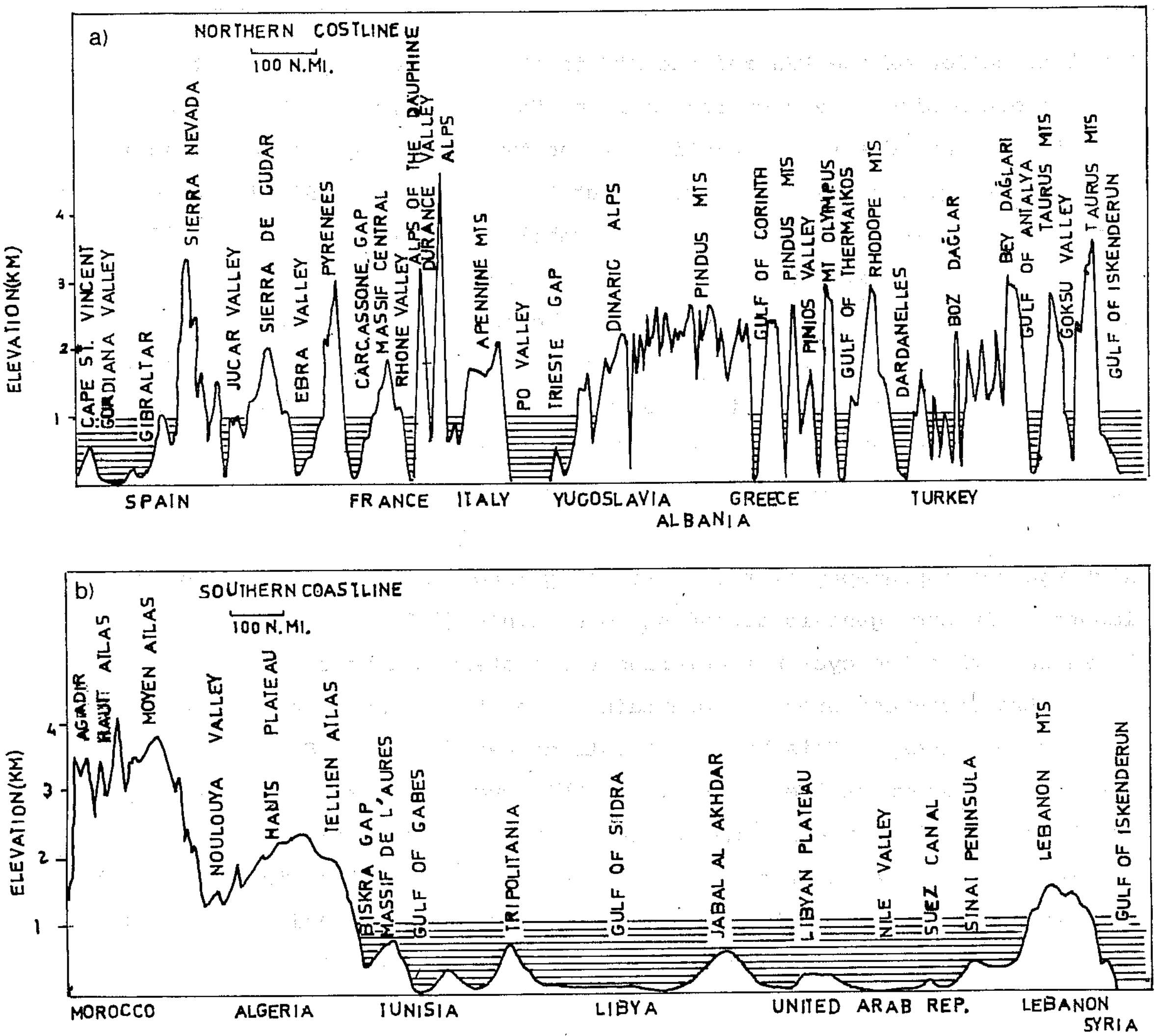


Fig. 7 Cross sections, approximately parallel to (a) the northern and (b) the southern shores of the Mediterranean, and roughly 100 miles inland from the shore. Major orographic obstacles and gaps between mountains are indicated. Gaps below 1000 m are shaded. (After Reiter, 1975).

very rapid trigger phase, the low-level front interacts with the Alps, and during the following baroclinic development phase a three-dimensionally coherent lee cyclone is produced when the low-level development merges with the larger scale upper-level disturbance. This may force one branch of the PFJ to place itself over the gap between the Pyrenees and the Alps. The other branch passes to the north of the Alps. (Trevisan, 1976, Buzzi and Tibaldi, 1978, Tibaldi, 1979).

The development is caused by a complex mechanism in which other factors, such as the geometry of the obstacle, also may play important parts. While the Genoa cyclogenesis is commonly associated with meso-scale wind systems such as the Mistral and the Bora, it has been suggested that the low-level cold-advection by Mistral further strengthens the cyclogenesis (Buzzi and Tibaldi, 1978).

The interaction of the PFJ and the STJ is shown to occur prior to the central and eastern Mediterranean cyclogenesis by Karein (1979), such that the upper-air wind maxima, due to the meeting of the two jets, occur as an essential precursor to cyclogenesis observed about 12 hours later. Sinking of cold domes also plays a role in the baroclinic instability. However, Karein (1979) concludes that the upper air development and a combination of barotropic and baroclinic instabilities is responsible for the eastern and central Mediterranean cyclogenesis. Neither the heating of cold outbreaks upon entering the Mediterranean nor the mountain blocking are sufficient reasons for these cyclonic developments, but the heat transfer from the sea to the atmosphere on some occasions gives rise to secondary cyclones.

Although the topography of the surrounding land masses does not seem to be important in cyclogenesis according to Karein's (1979) studies, it is a well known fact that the cyclones entering the eastern Mediterranean are steered by the two east-west oriented mountain ranges to the south and north of Asia Minor respectively. This branching into either the Black Sea or the Levantine sea has been shown earlier in Fig. 3. This suggests a wave guide effect of the Anatolian plateau which may be responsible for coastal trapping of atmospheric waves such as that studied by Gill (1977). The contributions of the terrain and the gaps located along the southern coast of Asia Minor to the Cyprus cyclogenesis also need to be studied in further detail.

Fig. 8 shows the distribution of mean sea-level pressure during the summer and winter seasons. In summer the land portion and especially the mountainous areas are characterized by low pressures, whereas the regions over the Mediterranean sea are generally high pressure areas. In winter this situation is reversed, with highs over land with maximum values reached in mountainous areas and lows over the Mediterranean Sea. This is mainly due to the differential heating of land and sea during the two seasons. Note also that the pressure gradients in winter are intensified along the northern borders of the Mediterranean, where the almost continuous mountain chain is located. During summer, the interior and the southern coasts of Asia Minor are influenced by the permanent low located in the Persian Gulf area. A high pressure ridge is located near the Balkan Peninsula. This summer situation gives rise to the strong and persistent northerly Etesian winds in the Aegean Sea (Reiter, 1975), whereas most of the remaining Mediterranean waters are subject to the mid-latitude westerlies. A north-westerly flow due to Mistral events is observed

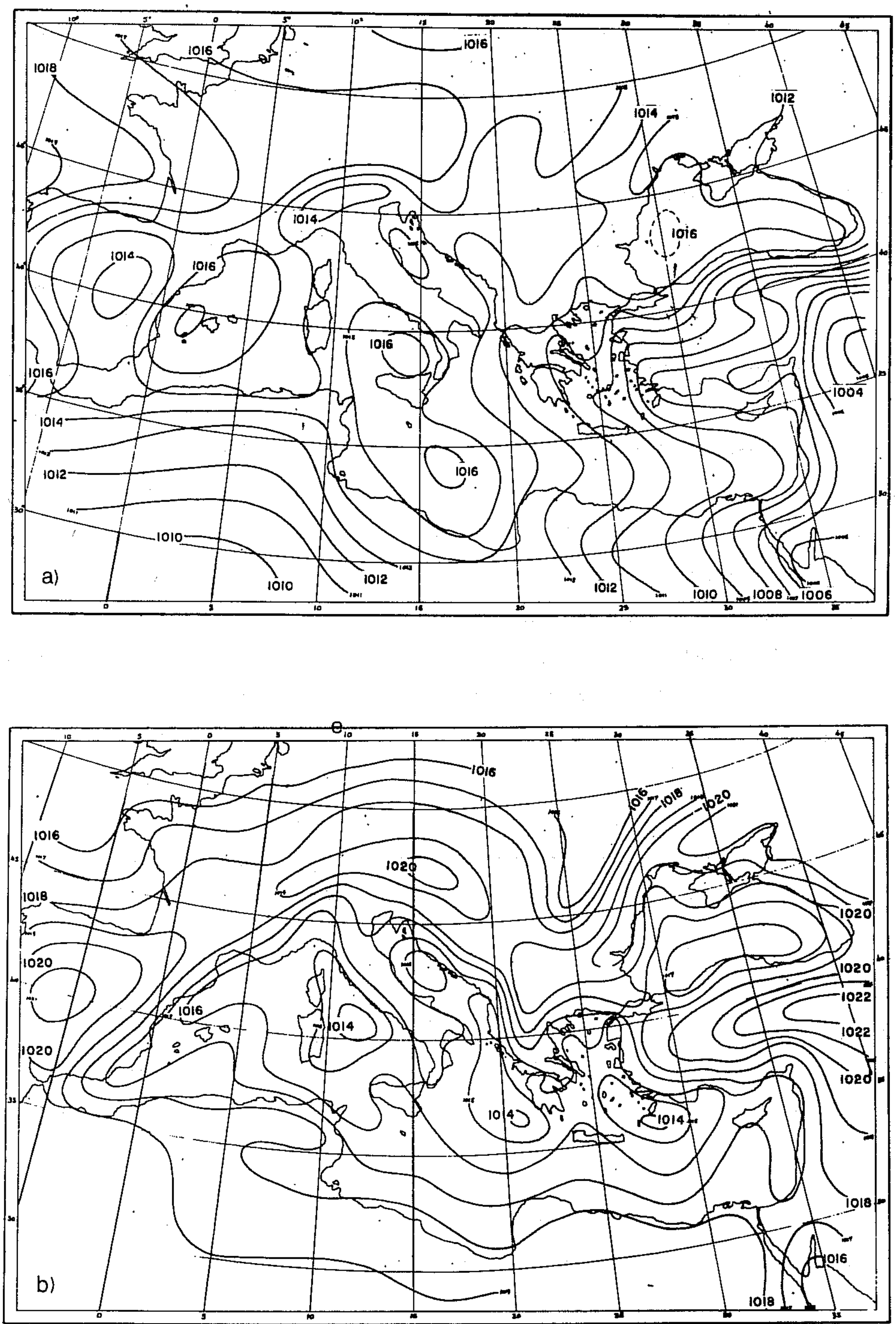


Fig. 8 Mean sea-level pressure (mb) during a) summer and b) winter. (After Reiter, 1975).

in the western Mediterranean even during the summer season (Middellandse Zee, 1957). In winter, low pressures are observed at the three cyclogenesis regions (Fig. 8). The high pressure area on Asia Minor is connected with the Siberian high during certain winter periods and highly stable cold and dry air fills the Anatolian Plateau behind the Taurus mountains.

3.2 Meso-scale phenomena

The role of mountain ranges and mountain gaps in the interaction between synoptic and meso-scale processes has been stressed earlier. Many local wind systems are generated within the Mediterranean region as a result of its complex topography (Fig. 9). In particular, it is well known that the blocking of a continuous low-level front by the Alps often gives rise to Mistral on its western flanks and Bora on the eastern flanks (Crowe, 1971, Buzzi and Tibaldi, 1978). Due to the channelling effects of gaps maximum hourly winds of 40 - 50 m/sec are observed in both regions. Trajectories computed by Buzzi and Tibaldi (1978) during a lee-cyclogenistic episode indicate (Fig. 10) descending motions in the lower atmospheric layers below the top level of the Alps and cold outbreaks in both the northern Ligurian Sea and the northern Adriatic areas. The trajectories on the eastern side make a turn around the Alps and towards the Po valley. The trajectories at higher levels indicate descending motions in the Mistral region behind the centre of the depression and strong ascending motions over Italy and the Alps, which is consistent with observations. Mistral winds often bring clear skies due to the descending motions of dry upper air. While descending Mistral winds extend to elevations of 2 - 3 km above sea level (Reiter, 1975), Bora is confined to lower layers with a reversed flow aloft (Clark, 1977). Mistral represents the channelling of descending motions behind Alpine lee depressions and is sensitive to low-level pressure patterns due to the rather large size of the gap geometry. Bora on the other hand is a much different phenomenon of truly katabatic origin (Reiter, 1975).

The accumulation of cold air over the Balkan Peninsula is a prerequisite for the occurrence of Bora. When this cold air pool reaches at least up to the mountain passes leading through the Dinaric Alps, a katabatic flow is triggered, where the cold air descends above the mountain slopes in the form of fall winds (Crowe, 1971). For this reason, not much air movement needs to occur to trigger these winds (Reiter, 1975). Anticyclonic Bora occurrence during summer is characterized by a high pressure system over Central Europe, but no well developed low to the south. Skies are generally clear and the wind does not extend more than 15 km offshore. Cyclonic Bora (Bora Scura) on the other hand occurs whenever a cold outbreak reaches north of the Mountain range or whenever

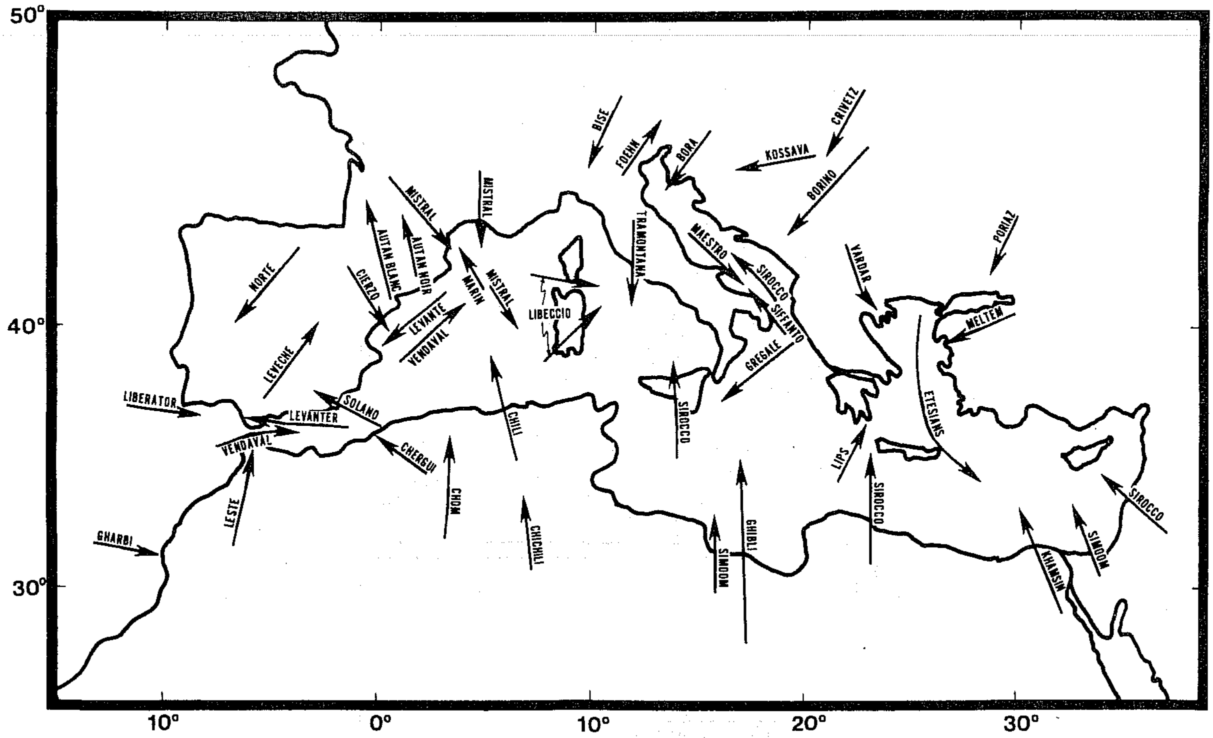


Fig. 9 Major local wind systems in the Mediterranean region. (After Reiter, 1975)

a depression over Central Mediterranean draws northerly winds over Yugoslavia. These winds are associated with a cloud cover with occasional rain and snow (hence the name Bora Scura: Dark Bora), and the winds cover the whole width of the Adriatic (Crowe, 1971, Reiter, 1975).

The meso-scale effects of mountain gaps are significant along the southern Turkish coast. The topography above the 1000 m level is shown in Fig. 11. There are mainly three major gaps in the mountains: at the Gulf of Antalya, the Göksu river valley and the Gulf of Iskenderun (see also Fig. 7). At the Gulf of Antalya and the Gulf of Iskenderun regions, changes in the orientation of the Taurus mountains allow a connection with the interior. The only other passage is through the Göksu river valley and this "Riviera" coastland is connected to the interior by roads passing through these three gaps. Özsoy et al (1980) have shown that the most violent winds along the Turkish coast occur at the Göksu valley region. The descent from the interior to the coastland in fact occurs in two steps. The mountain pass through which the Goksu river flows is at about 1500 m elevation (which is higher than the 1000 m high interior Anatolian Plateau), then descends into the Mut basin which has a rather large diameter of 30 - 40 m and a mean elevation of 300 m. A narrow valley then threads down from the Mut valley to the Goksu river delta near Silifke. The Mut valley is separated from the coastland on the western side by the 1500 m high Taseli plateau. This channelling geometry gives rise to the Poyraz winds which bear much similarity to the Bora.

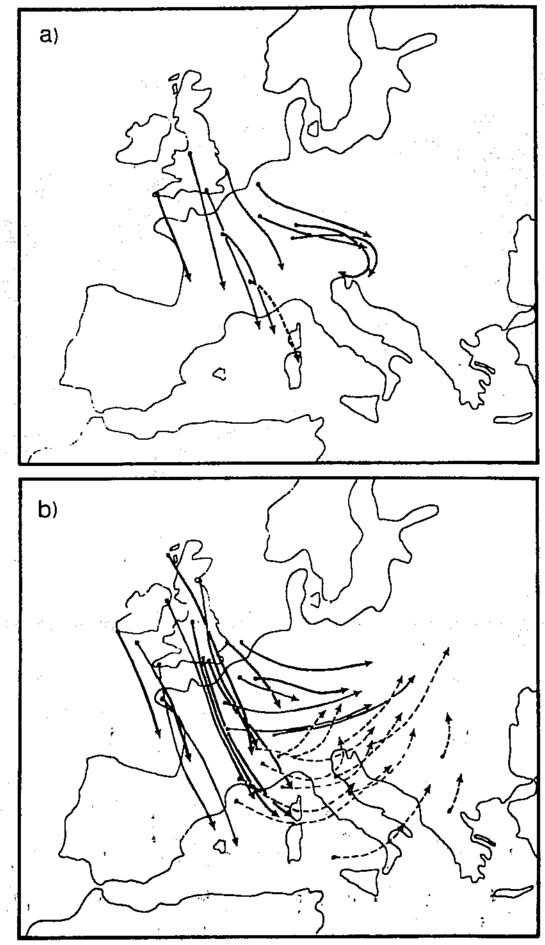


Fig. 10 Isentropic trajectories: 3 April 1973, from 00 to 12 GMT. Full lines: descending trajectories; dashed lines: ascending trajectories. (a) θ = 285 K. (b) θ = 295 K. (After Buzzi and Tibaldi, 1978).

Time series for wind stress (x and y axes directed towards the east and north, respectively) at Akkuyu (located to the south of the Taseli Plateau), relative humidity at Silifke, and sea water temperature at Akkuyu are plotted in Fig. 12. The strong northerly winds observed at Akkuyu mark the Poyraz events, superimposed on rather uniform sea-breeze oscillations of diurnal period. The monthly mean wind stress at four coastal locations in Fig. 13 shows that the Poyraz is effective in an area extending from Anamur to Silifke (a distance of nearly 100 km) especially during the winter season and during some summer months (mainly August).

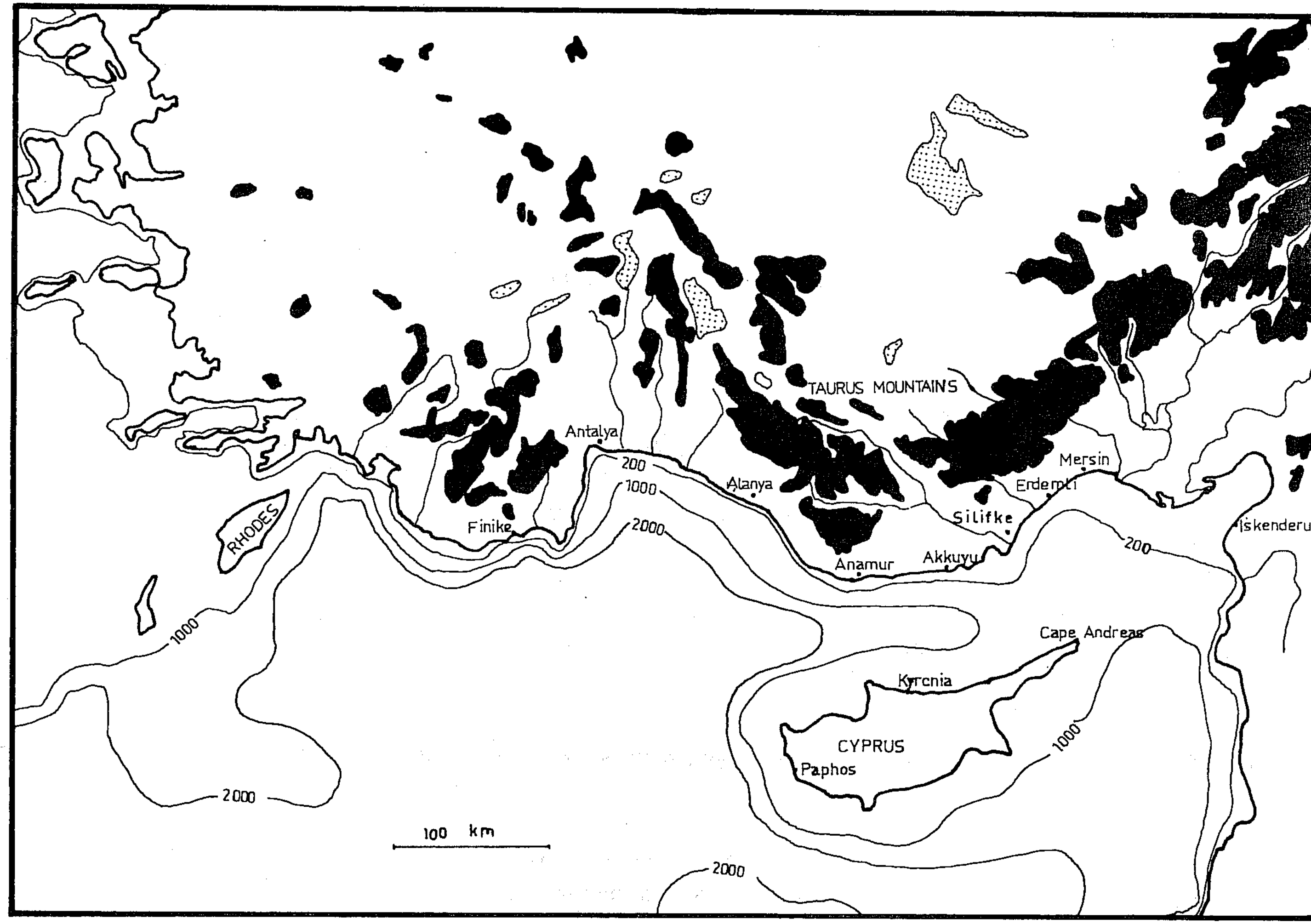


Fig. 11 Topography of the southern Turkish coast. The shaded areas have elevations above 1000 m.

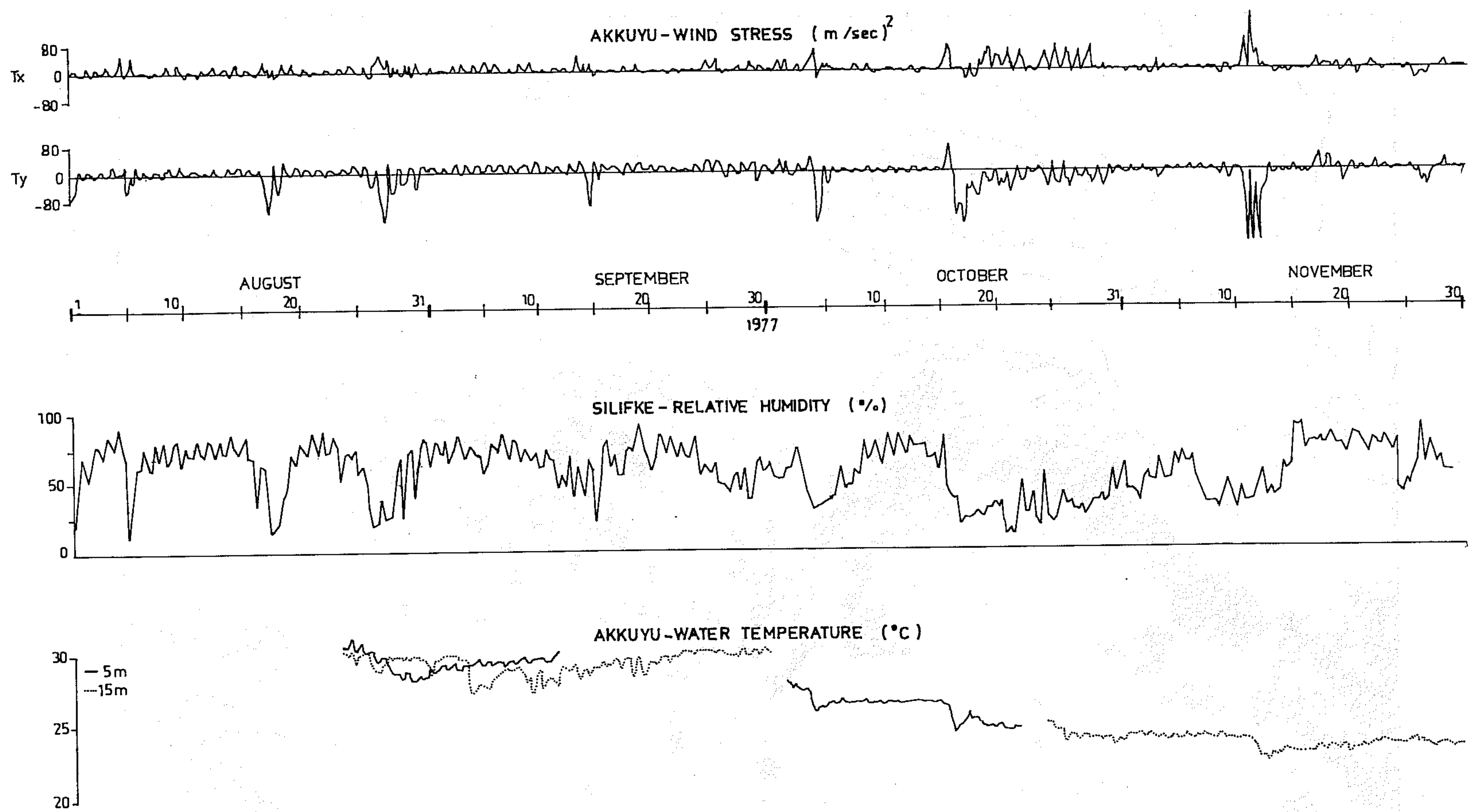


Fig. 12 Time series of wind stress, relative humidity and sea water temperature during four months in 1977.

During August, no cyclones are observed to the south of the valley, and Poyraz winds occur when the temperature difference between the interior and the coastland is increased (not shown in Fig. 13) either due to a cooling in the interior due to depressions passing to the north of Asia Minor or by warm air masses arriving at the coastal area (Ataktürk, 1980). This purely katabatic flow case is similar to the anticyclonic Bora. In the winter season, frequent occurrences of Poyraz are triggered, with depressions passing along the southern coast of Turkey or across the Anatolian plateau (only a four-month period is shown in Fig. 13, the occurrence frequency is increased in the following months of winter and early spring). During each Poyraz event, extremely dry air (relative humidity falling down to 6% in the observed period of Fig. 13 but with minimum of 4% observed on occasions) is brought to the coastal area. In summer months, the dry air comes from the upper levels of the air mass pooled in the interior. In winter months, the descending motions behind passing depressions are channeled through the valley and bring dry and cold air from the upper air layers (such as in the case of Mistral).

It is well known along the southern Turkish coast that even in winter the passing depressions are usually followed by bright skies and dry conditions which suddenly clear the dense clouds observed earlier. Especially after Poyraz events the visibility is so much improved in the Akkuyu-Silifke area that the mountains of Cyprus are visible from a distance of 100 km or more.

The seaward extent of the Poyraz winds is not well-known. Based on limited data available at Kyrenia, Cyprus, Ataktürk (1980) indicates that northerly winds with much reduced velocities are observed on Cyprus during Poyraz. Based on reports of boats lost on the Turkish coast and found at Cyprus, and ship observations it can be concluded that these winds reach great distances offshore during winter. There is a significant shear line especially near Silifke, where winds suddenly cease to the east of this region (see Fig. 13).

Ataktürk (1980) has compared rotary spectra of wind stress at the four stations along the southern Turkish coast. These indicate that Poyraz winds affect a region from Anamur to Silifke. The diurnal sea-breeze oscillations are quite strong along the whole eastern Mediterranean coasts and the coasts of Cyprus. The sea-breeze has a clockwise rotation due to Coriolis forces at all stations except at Silifke which is located near the Göksu river valley mouth.

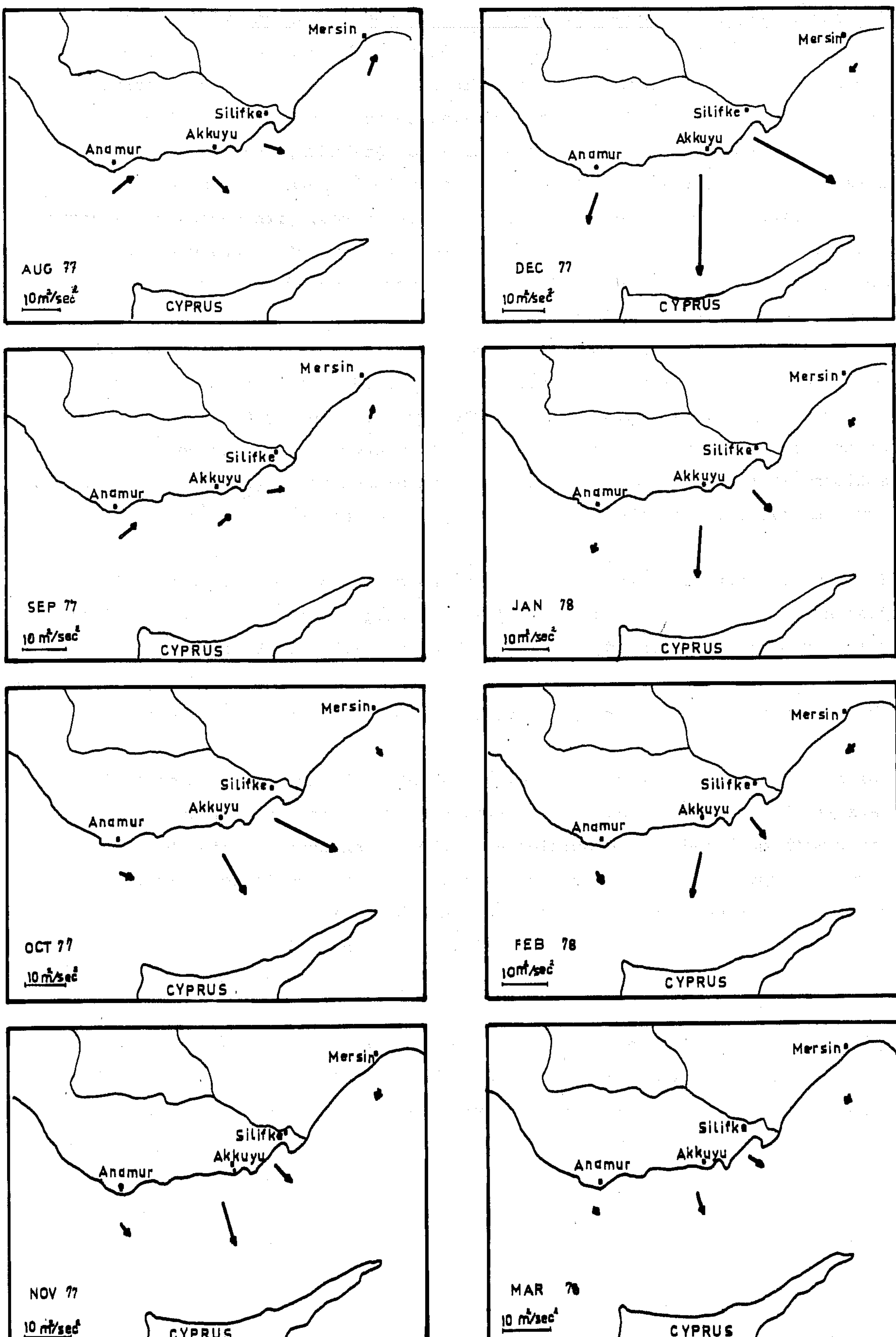


Fig. 13 Mean wind stress along the southern Turkish coast

In order to demonstrate the effects of synoptic-scale processes on the local Poyraz winds, weather maps corresponding to the events starting on October 16, 1977 and November 11, 1977 (see Fig. 12) can be studied. On the event of October 16 a southwesterly wind precedes the northerly winds since during this period a cyclone moves along the southern coast of Turkey and crosses the area (Fig. 14) first inducing southwesterly winds due the approaching cyclone centre, then triggering the northwesterly Poyraz winds. As opposed to this case, strong northwesterly winds during the Poyraz on November 11 do not have preceding southwesterly winds, since in this case a cold outbreak reaches the area from the north.

The October 16, 1977 Poyraz winds are followed by a long period of strong northwesterly winds with diurnal variability. This long duration of strong continental winds did not occur during the other periods studied between August 1977 and June 1978 (Ataktürk, 1980). The development leading to the formation of the cyclone passing along the Mediterranean is shown in Fig. 15. A cutoff low develops over the Adriatic sea on the 12th together with a ridge formed over Spain (this ridge and the cutoff low are also clearly observed at the 200 mb level, not shown here). On the 13th this cold dome over Italy generates a surface cyclone over the central Mediterranean, while the high pressure ridge advances into central Europe. On the 15th the surface low reaches the western shores of Turkey, while the high pressure ridge develops into a cutoff high in central Europe. This blocking high occupies central Europe until November and affects the weather during this period. It is seen to be present at all pressure levels. Although the blocking high oscillates around a mean position, plots of successive positions of surface fronts during the periods 12-23 October and 24 October-1 November (Fig. 16) show the presence of the blocking high over Central Europe. The surface blocking high occupies the Balkan Peninsula and creates pressure gradients perpendicular to the Taurus Mountains, and since it prevents westerly flow the northerly winds cause ponding of cold air in the interior region. When the cyclone passage triggers the Poyraz winds, the favourable stationary pattern force the flow to continue for a long period. The above phenomenon is a good example of the influence of synoptic scale processes on the persistence of meso-scale local winds.

Before the event of November 11, 1977 a blocking high still appears over the Balkans. Successive positions of fronts starting from 6 November (only 5 days after the last data in Fig. 16) are shown in Fig. 17. Due to the blocking of the stationary high a highly meridional flow occurs again with a cold outbreak reaching Turkey from the north. The advance of the surface front is very rapid; the frontal perturbation crosses the Anatolian Plateau from north to south in 24 hours (Fig. 18) and causes the strong Poyraz winds observed during this period.

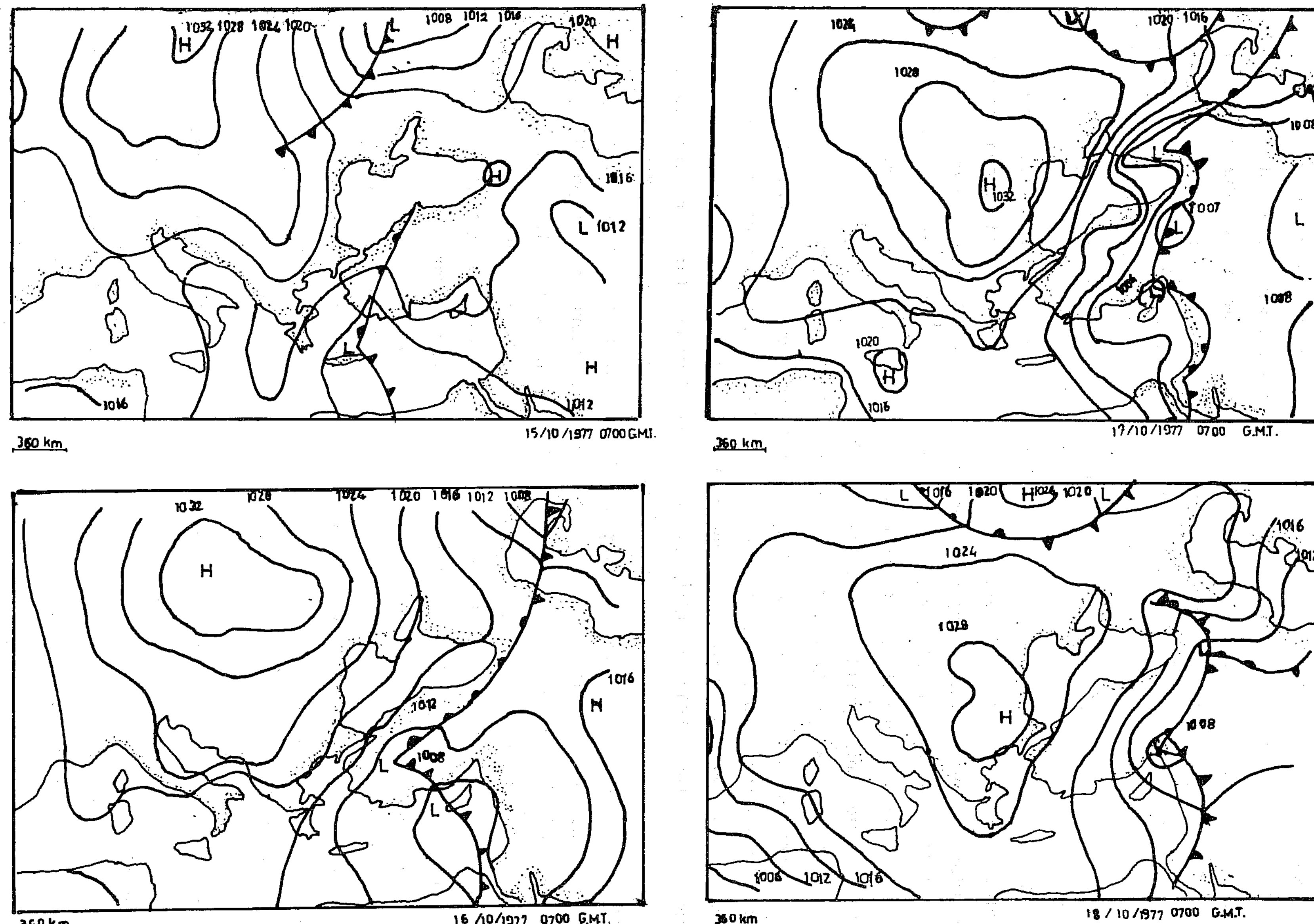
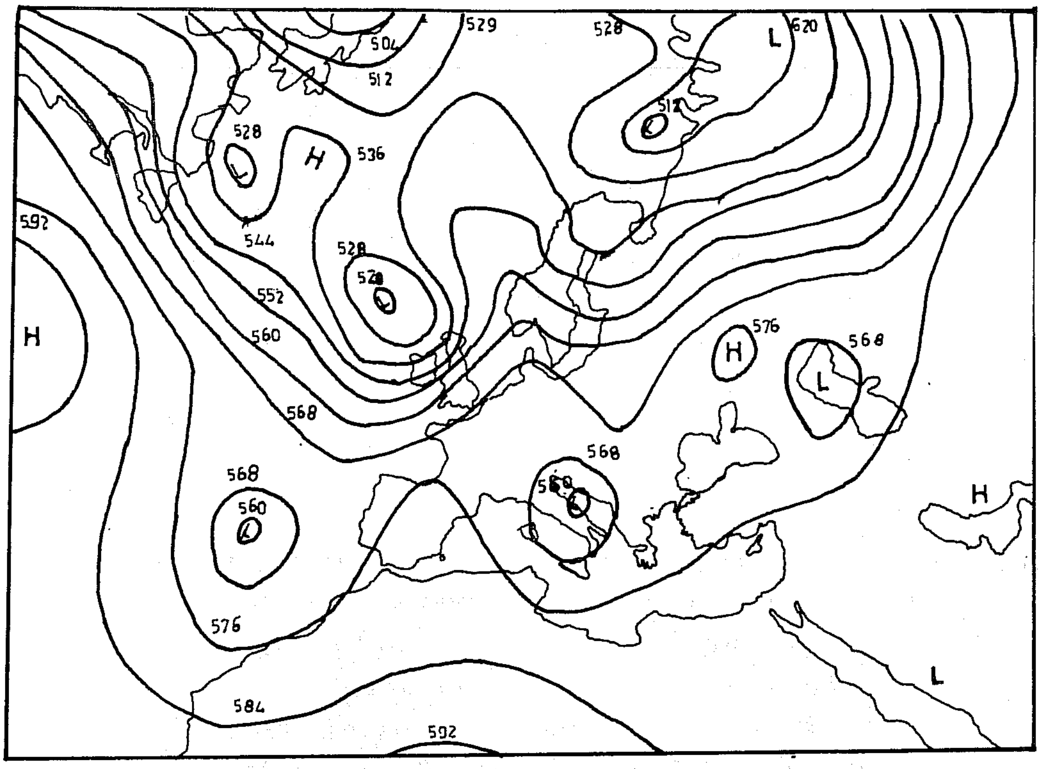


Fig. 14 Synoptic surface maps, October 15-18, 1977. (Turkish Meteorological Service Maps).



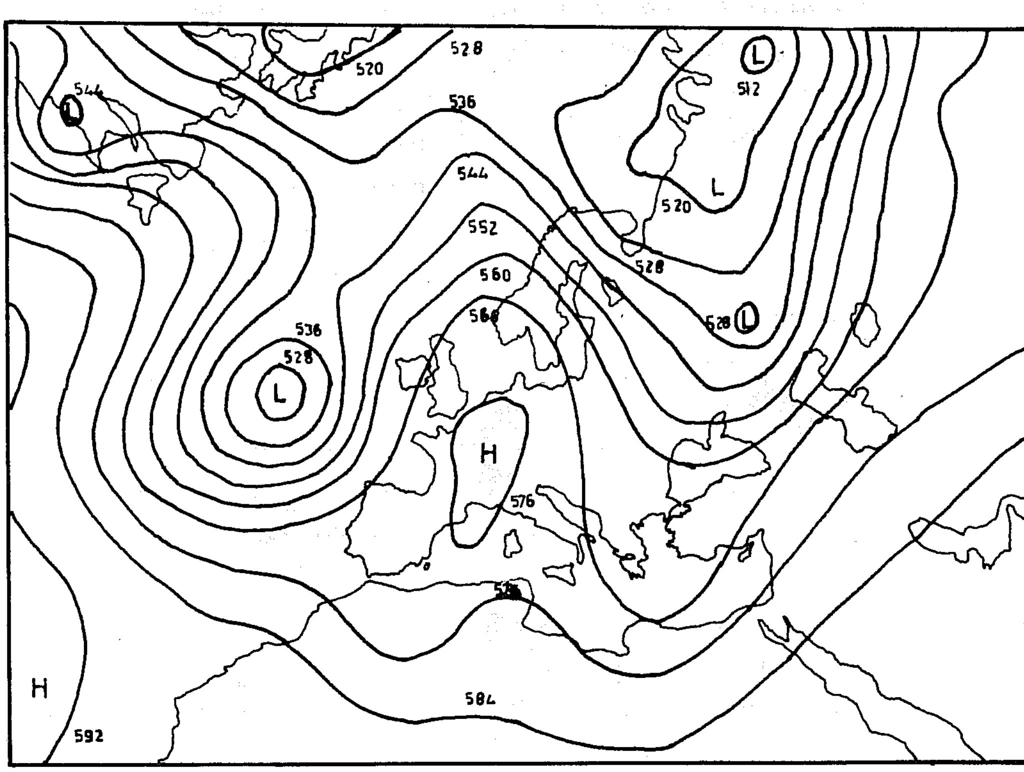


Fig. 15 500 mb Analysis a) 12 Oct 1977 00 GMT (Source: European Meteorological Bulletin, Deutscher Wetterdienst).

500 mb Analysis b) 15 Oct 1977 00 GMT (Source: European Meteorological Bulletin, Deutscher Wetterdienst).

Fig. 16a Successive Positions of Surface Fronts, 12-23 Oct, 1977 (compiled from European Meteorological Bulletin, Deutscher Wetterdienst).

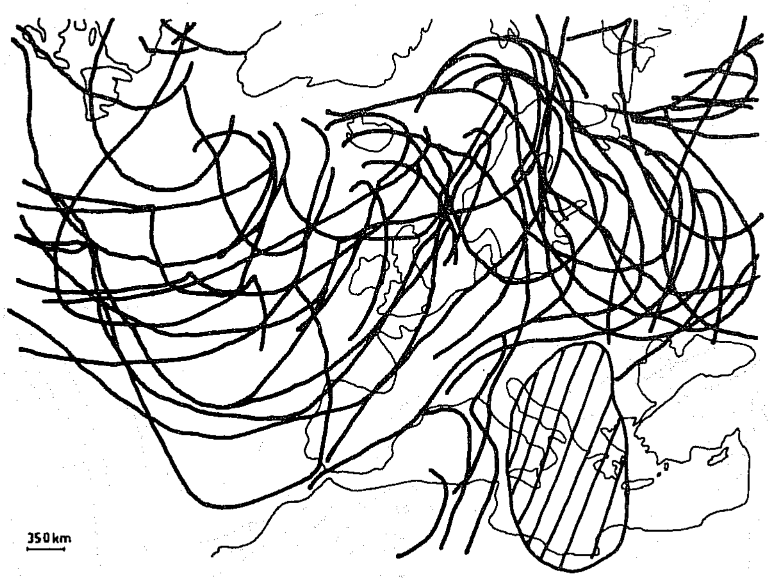


Fig. 16b As for Fig. 16a. 24 Oct-1 Nov 1977.

Fig. 17 As for Fig. 16a. 6-12 Nov 1977.

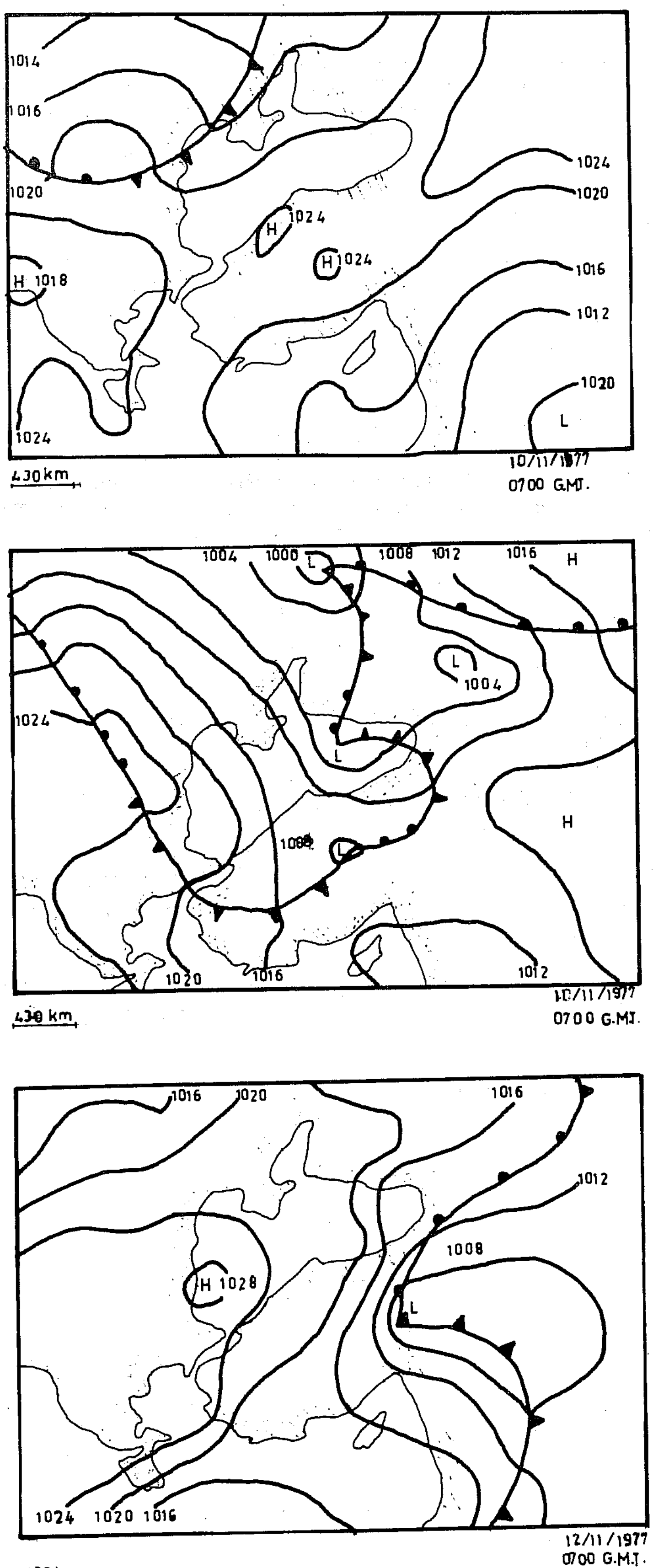


Fig. 18 Synoptic surface maps, 10-12 November 1979
(Turkish Meteorological Service Maps)

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