DYNAMICAL ASPECTS OF THE CILICIAN BASIN-
NORTHEASTERN MEDITERRANEAN

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1. INTRODUCTION

The insufficiency of the data base is one of the major reasons why the eastern Mediterranean system is the least understood among the various Mediterranean areas. In this region, the fundamentals of descriptive, kinematical and dynamical processes are often controversial and ill-defined, although a better understanding of these processes is of crucial importance in climate prediction, utilization of resources in fisheries, energy, transportation, industry and tourism sectors and in resolving pollution aspects (Round Table of the Eastern Mediterranean, c 1982)

We review dynamical aspects in the northeastern Mediterranean with special reference to the Cilician basin (area between Turkey and Cyprus). Through a number of studies carried out recently by the Institute of Marine Sciences, Middle East Technical University, a better than nonexisting knowledge has been acquired on this portion of the eastern Mediterranean. We first report on the atmospheric variability in section 2. Mean circulation is reviewed in section 4. A review of observations on the low and high frequency motions is given in sections 5 and 6. A discussion on water masses follows in section 7 and recommendations are provided in section 8.

2. THE ATMOSPHERIC SETTING

A review of the atmospheric setting is essential in leading the discussion on dynamical characteristics of the region and in bringing into perspective the climatological peculiarities of the Mediterranean basin.

The prevailing meteorological conditions in the north-eastern Mediterranean are extremely variable. In winter and spring the area is a pathway for extratropical cylones (Fig.1), some 30 depressions passing along the southern coast of Turkey annually (Hydrographer of the Navy, 1976; Karein, 1979). In summer and autumn, the mid-latitude westerlies are modified by a strong sea-breeze regime. The coastal topography supports localized wind regimes, which introduce additional variability (Özsoy, 1981).

These regional features have much in common with the rest of the Mediterranean, a unique area of world in terms of the many complex atmospheric processes created by influences of topography, land and water mass configurations. A comprehensive review by Reiter (1975) suggests that the planetary, synoptic, meso - and micro-scales of motion interact and often act in unison. A review with special reference to the Levantine basin is given by Özsoy (1981).

Of the 76 Mediterranean cyclones observed annually (Fig.1), 69 are locally generated (Karein, 1979) often due to Alpine lee cyclogenesis (Buzzi and Tibaldi, 1978; Tibaldi, 1979) or interaction of atmospheric jets and heat transfer from the sea (Reiter, 1975; Karein, 1979). Regardless of the instability mechanisms, the semi-permanent Mediterranean front (Peterssen, 1956) forms a suitable environment for cyclogenesis, enhanced by topography (Reiter 1975).

While the Alpine topographic obstacle increases contrasts between Europe and the Mediterranean, numerous gaps in the form of river valleys and mountain passes (Fig.2) allow low level communication through them, resulting in local wind regimes such as the Mistral and the Bora (Reiter, 1975). These local winds are often triggered or strengthened by cyclones or cold outbreaks in the cold season, and winds of purely katabatic origin blow through the same gaps in summer (Crowe,1971; Bunker, 1972; Buzzi and Tibaldi, 1978; Reiter, 1975; Brody and Nestor 1980).

A similar situation is created on the southern Turkish coast by major crossings of the Taurus Mountain range at the Gulf of Antalya, the Göksu river valley and the Gulf of Iskenderun regions (Fig.2), which allow locally strengthened wind regimes by connecting the interior to the riviera coastland (Fig.3). Among these three passages the Göksu river valley has the most violent wind regime blowing from the NW, which is locally known as Poyraz (Özsoy et al,1981).

The characteristics of the Poyraz wind regime are reflected in Fig.4 (Özsoy, 1981). The small diurnal fluctuations in wind stress (in units of m^2/s^2) represent the sea-breeze regime, the single events with strong northwesterly winds being due to the Poyraz regime. The meso-scale Poyraz winds are of katabatic origin during summer and are often related to cyclonic passages and synoptic-scale processes in winter (Ataktürk, 1980; Özsoy, 1981). Each Poyraz event pumps extremely dry upper atmospheric air to the coastal region (minimum observed relative humidity 2%, compared to annual average of 65%), and maximum hourly winds reach up to 36 m/s (based on 1976-79 statistics, Tamer et al, 1980). The resulting mixing and evaporative transfer from the sea surface causes rapid, stepwise cooling of surface waters as displayed in Fig.4.

An important consequence of the local wind regimes rests in the spatial variability of the wind-stress field. The Poyraz regime is effective in the region extending from Anamur to Silifke (Fig.3) as indicated by the monthly average wind stresses of Fig.5. Although the seaward extent of Poyraz is not well known, Ataktürk (1980) indicates that the island of Cyprus is influenced during winter months.

Rotary spectral analyses of wind-stress at four coastal stations are given by Ataktürk (1980) on a monthly basis. Representative spectra for the months of August and November 1977 shown in Fig. 6 indicate the variability of winds along the coast. In August, diurnal clockwise-rotating (excluding Silifke) sea-breezes dominate the coast and low frequency spectral components at Akkuyu and Silifke reflect Poyraz events. Again in November, low frequency energy appears only at the three stations excluding Mersin, a station outside the Poyraz regime.

3. BOTTOM TOPOGRAPHY

The bottom topography of the northeastern Mediterranean is shown in Fig.7. The Turkish coast is characterized by a very narrow (steep) continental shelf which gradually widens towards the Gulf of Iskenderun on the east. Deep basins of Antalya (2000m) and Rhodes (4000m) are located on the west. The shallower Cilician basin (1000m) occupies the area between Cyprus and Turkey, connected to another 1000 m. deep basin to the east of Cyprus.

4. MEAN CIRCULATION

A cyclonic circulation in the eastern Mediterranean has been proposed as the dominant mean current system (Wüst,1961; Lacombe and Tchernia, 1972). Accordingly, the steady surface current follows the coasts of Israel, Lebanon and Syria and turns west to flow along the southern Turkish coast. Using the observed density and wind fields, Ovchinnikov (1966), Engel (1967), Moskalenko (1974) and Gerges (1976) have confirmed computationally this picture of mean motion. Collins and Banner (1979) have utilized the computed geostrophic fields together with ERTS imagery and secchi depth measurements to provide the details of the flow in the northeastern corner of the Levantine basin, shown in Fig.8. The patterns of the subsurface flow essentially follow that of the surface motion (Gerges, 1976).

The existence of a mean westerly current along the southern Turkish coast has been confirmed by recent observations (Ünlüata et al, 1978,1980,1983). Results of all mean current measurements after 1977 (obtained from current-meter deployments at sites E and A, Fig.9) on the Turkish continental shelf are summarized in Tables 1 a, b (Ünlüata et al, 1983). While the westerly mean flow is detectable with an order of magnitude of 10 cm/s at site E located along the relatively smooth coastline extending from Mersin to the Göksu river (Fig.9), the current magnitude is significantly reduced with no predominant direction at site A and in the nearshore areas west of the Göksu delta (Table 1.b). This is even more clearly demonstrated by the simultaneous measurements of summer 1980, since a mean current of 14 cm/s is observed at site E as compared to the small currents obtained from two current-meters at site A (Tables 1 a,b).

The reduction of mean westerly currents near the rugged coastline to the west of the Göksu delta led Ünlüata et.al (1983) to investigate the blocking action by coastal geometry. Topographic steering by the bathymetry of the Göksu delta and the continental shelf, rather than flow separation in the lee of headlands, was found to be responsible for the observed blocking of the mean circulation. This result is also confirmed by numerical solutions as shown in Fig.10 (Ünlüata et al.1983). The effects of bottom friction were found to be small in deflecting the topographically steered currents.

5. LOW FREQUENCY MOTIONS

A well known fact about the eastern Mediterranean in general, and the northeastern Mediterranean in particular, is the insufficient understanding of the dynamical processes in these waters. Recent experimental data collected through current-meter deployments (Table 1) have yielded valuable (although far from complete) information on the dynamics of the Cilician basin. The outstanding result of these observations is the dynamism displayed by strong reversing currents (Ünlüata et al, 1978, 1980, 1982) which distinguish the region from the rest of the east Mediterranean (Hydrographer of the Navy, 1976). The striking space-time variability of the regional winds outlined in section 2 supports these findings.

The persistence of strong fluctuating currents is reflected by the standard deviations of current components in Table 1. The current components at site E (Erdemli) have been rotated by 45^{o} to align the u_R component along the straight coastline extending in the NE direction (Fig.9), which results in larger along-shelf standard deviation. The currents at site A (Akkuyu) are more disorganized but they also follow local bathymetry as to be shown later. At both sites the fluctuating currents overwhelm the mean components. The observed fluctuating currents are a superposition of low and high frequency components which are separated by filtering at cutoff periods near 24 hours.

A long-term (5 month) experiment was carried out at site E (Erdemli) during 26 August 1978 - 31 January 1979. The along-shelf currents were mainly long period fluctuations, whereas the cross-shelf flows were very small and had only high frequency (above diurnal) components. The low pass filtered along-shelf current component is shown in Fig. 11. The corresponding raw and smoothed spectra obtained through digital fourier transforms (DFT) and the maximum entropy method (MEM) are shown in Fig. 12. A very sharp peak in the low-frequency spectra occurs at 16 d, followed by 40 d, 10 d, 6 d, and 4 d periods.

Along-shelf currents and local-wind stress variations during the experiment do not appear to be strongly correlated in the low frequencies. Considering the non-uniformity of the wind field in the Cilician basin (section 2), it is not unusual to have lack of immediate correlation with local winds, although the distribution of regional winds as a forcing agency should be expected to play an important role in exciting the low frequency motions.

Based on the indirect effects of local winds in driving the continental shelf motions along with the sharpness of some peaks in the spectra of Fig. 12, Ünlüata (1982) proposed a resonant mechanism. Through an analysis of motions forced by various wind stress distributions it was found that the motions including those with periods of 16 d. and above could be manifestations on the continental shelf of resonant topographic Rossby waves in the interior of the Cilician basin. Bottom topography in the interior of the basin deepens from Cyprus towards Turkey, where it meets a stepp continental shelf (Ünlüata, 1982). The opposing slopes of the interior and the shelf regions does not allow phase-matched resonance of the shelf, although forcing by wind stress at cutoff frequencies (i.e. yielding zero group velocity) of the topographic waves resonates the interior region and thus can be detected on the shelf. Ünlüata (1982) calculates a succession of resonant periods at 16 d, 31 d, 47 d, etc., some of which are confirmed in Fig 12, but the succession towards low frequencies with decreasing intervals puts a large requirement on data length which can hardly be satisfied by even the five month period of Fig.11.The peaks in the spectrum of Fig. 12 corresponding to periods of 3-10 days can not be explained by topographic waves of the interior, and are possibly due to Kelvin and shelf wave modes of the shelf region.

A separate experiment was carried out during June-September 1980 in which current-meter data were obtained simultaneously at sites E and A of Fig. 9, separated by distance of 90 km. Two current meters were placed at Akkuyu with horizontal separation of 500 m (stations A1 and A2). The low-passed time series (at 30h) of currents and temperature are shown in Fig 13. The currents at stations E1 and A1, A2 have been rotated by 45° and -40° respectively to align u_R components along the local bathymetry. The low-frequency currents at site E are of larger amplitude and follow the bathymetry more closely as compared to site A, where the coastline is more irregular. The low-passed temperature records show large amplitude cooling periods which mainly correspond to westerly flow conditions (marked by the vertical lines in Fig. 13), superposed on a seasonal warming trend. Currents at both sites can be visually correlated and show similar patterns of reversal. Both the large cooling events and current reversals have 14-17 day periods as marked in Fig. 13. A nonconforming peak occurs in station E1 currents and temperature around July 22, and the last two events in the temperature series do not appear to be related to westerly flow. The 14-17 day periodicity in the low-passed data is important in that it shows time-dependent upwelling related to the 16 day lowest mode resonant topographic waves of the Cilician basin modeled by Ünlüata (1982). Since an inspection of the local wind-stress does not indicate a strong correlation with currents, the corresponding time series is not shown.

The very low frequency (periods above 5 days) and low frequency (1-5 day range) variability is better demonstrated by low (Fig. 14) and high-pass (Fig. 15) filtering the time series of Fig. 13, at 5 day cutoff period, after removing trends from the temperature series.

The shortness of data length as compared to the low frequency quasi-periodic signals restricts the quality of spectral analyses, which usually can not be overcome by smoothing. As a first attempt, spectral analyses have been made on the full lengths of the low frequency time-series, using digital fourier transforms (DFT), after removing averages and detrending the temperature time-series (Fig. 16). Smoothing by three-point moving averages (solid lines) increases stability at the cost of resolution, but persistent peaks such as the 14-18 d, 8-10 d, 6 d, 5 d, 3 d are detected in the current and temperature spectra. The wind stress spectra for Akkuyu is full in low frequencies with peaks at 20 d, 13 d, 9 d, 6 d periods. To improve the results further, maximum entropy method (MEM) has been used to estimate current spectra (Childers, 1978; Marmorino, 1978), yielding better resolution (Fig. 17). The spectral peaks corresponding to 16 d, 7 d, 6 d, 4.5 d and 3.5 d periods are detected in all three stations. The 10 d peak is detected in E1, A1 currents, but there is somehow no evidence of this period at station A2.

In summary, recent observations reflect orderly low frequency motions on the continental shelf which can partly be explained in terms of the resonant barotropic response of the Cilician basin interior (periods > 16 days) and partly as possible wave modes of the shelf region (periods 1-10 days). Although immediate effects of local winds in driving these motions are not easily discernible, the presence of low frequency wind variability suggests the importance of wind forcing. It is usually accepted that the narrowness of shelf regions as compared to a more or less uniform wind stress field is the basic reason why shelf motions are uncoupled from the ocean interior (Allen, 1980). The non-uniformity of the wind stress field and bottom topography in the region along with the omnipresence of the interior currents on the shelf suggests possible interior-shelf coupling for some forced modes. Further investigations and detailed models including generation and instability mechanisms are needed to understand the nature of low frequency motions in the northeastern Mediterranean.

6. HIGH FREQUENCY CURRENTS

High frequency (periods 1 day and smaller) motions are evident in most current-meter records obtained during the deployment periods of Table 1. The most significant components are diurnal (24 h) and inertial (local inertial period being 20.4 h at 36° latitude) fluctuations. High-pass filtered data corresponding to the June-September 1980 period (Fig.18) will be used to outline high frequency characteristics.

Sea-breezes of diurnal period are dominant in the summer wind-stress time-series of Fig. 18. Sea-breezes show variation (see also section 2) from Mersin to Akkuyu (in 150 km), with diurnal components of Poyraz events also being present at Akkuyu. The high frequency currents and temperature at Akkuyu (A1, A2) are larger as compared to Erdemli (E1) since the sea-breeze system is generally stronger at Akkuyu. It can also be observed that the high frequency currents and temperature respond better to long period modulation of sea-breeze than single wind events.

Rotary spectra for the high-passed time series of Akkuyu wind stress and E1, A1, A2, currents, shown in Fig. 19 were obtained by averaging spectra from two (partially overlapping) 1024 h segments. Diurnal, inertial, semi-diurnal and 8 h oscillations are present in the wind-stress spectrum with clockwise sense of rotation. The same principal frequency components are found in current spectra of all stations, with mainly clockwise rotation. The only exception is the counter-clockwise rotating diurnal currents at station A1, notwithstanding its closeness to station A2. Inertial peaks in current spectra are significantly large when compared with the ratio of inertial/diurnal peaks in the wind-stress spectrum, most evidently at station E1. While inertial and diurnal current oscillations are dominant as a result of wind forcing, the semi-diurnal currents are most probably related to the semi-diurnal tides in the area. In addition, peaks at 30 h, 16 h, 10 h and 8 h periods occur in the current spectra. The origins of these peaks are not clear, although it is possible to explain such periods as harmonics of the 4 d (section 5), 2 d (not shown), 24 h and 20.4 h oscillations existing in the current spectra.

The closeness of diurnal and inertial frequencies in the region (separation 0.0074 cph) sets out a major problem in obtaining sufficient spectral resolution (0.002 cph resolution bandwidth in Fig. 19). In addition, the closeness of the diurnal frequency to the inertial (a principal natural mode of the ocean) makes possible the near-resonant excitation of inertial motions through diurnal sea-breeze forcing in the region. However, the observed transient nature of the inertial motions are often viewed as being incompatible with the resonance phenomenon, since turbulent frictional losses can not account for observed decay rates. The more subtle processes of radiation of energy, wavelength dispersion and transient wind forcing are responsible for the intermittency of inertial motions (Pollard and Millard, 1970; Pollard 1970; Smith 1973). In addition, geostrophic adjustment processes in relation to short term upwelling events near a coast can contribute to the evolution of inertial motions (Tang, 1979; Millot and Crepon, 1981).

Along the Turkish coast the sea-breeze system is mainly responsible for the near-resonant forcing of inertial motions in addition to driving the diurnal motions at the surface. In order to study the evolution of these motions, the time series of Fig. 18 have been been band pass filtered, (the results for station A2 displayed in Fig. 20). Episodes of inertial and diurnal motions occur in the corresponding bands (B1 and B2 respectively), which when combined (B3) make up for most part of the high frequency fluctuations observed in currents and temperatures. Comparing with Fig. 18 it can be seen that single Poyraz wind events have no effect on the generation of these motions, although they tend to destroy the existing motions. The modulations in the sea-breeze system appear to be better correlated with the high frequency motions at all stations in Fig. 18. It may also be noted that the inertial and diurnal episodes in currents do not appear at the same time with temperature. This may be due to the changes in stratification associated with the upwelling events of section 5.

7. WATER MASSES

The role of the eastern Mediterranean basin as a crucible molding Mediterranean water masses is well known.In much the same way as the greater Mediterranean Sea, three different water masses are found at the surface, intermediate and deep layers of the eastern basin (Nielsen, 1912; Wüst, 1961;Miller et al, 1970). However, due to inhomogeneities that are typical of source regions, it has not always been possible to trace these water masses to their origins within the eastern basin. For example, the Levantine Intermediate Water (LIW) which originates in the east becomes relatively uniform only after reaching the Ionian basin (Morcos, 1972). While investigations on the formation, circulation and budgets of water masses (Bethoux, 1980; Carter, 1956; Lacombe and Tchernia, 1960, 1974; Lacombe et al,1958; Miller, 1963, 1972; Morcos, 1972; Moskalenko and Ovchinnikov, 1965; Ovchinnikov, 1974, 1976; Pollak, 1951; Wüst, 1959, 1961) and on the role of climatological surface fluxes (Bethoux, 1979; Bunker, 1972; Bunker et al, 1971, 1982; Colacino and Dell'Osso, 1975, 1977; Lacombe, 1977; May, 1982; Morcos, 1972; Peixoto et al, 1982; Tixeront, 1970) are well documented on a Mediterranean scale, still very little is known today on the actual meso-scale mechanisms that may be responsible for the mixing and transformation that lead to water mass formation. Perhaps an approach that is technically similar in this context to the MEDOC compaign in the northwestern Mediteranean (e.g. Bunker, 1972; Gascard, 1978; Hogg, 1973; Killworth 1976, 1979; MEDOC Group, 1970; Sankey, 1973; Stommel, 1979) is much needed to identify detailed mechanisms.

It is generally accepted that the deep waters of the eastern Mediterranean are formed by Adriatic outflows (Pollak, 1951; Wüst, 1961; Zore-Armanda, 1977), although indications exist for some deep water formation in the Sea of Crete (Nielsen, 1912; Miller, 1963) or locally in the northeastern Mediterranean (Plakhin, 1973).

The Levantine Intermediate Water (LIW) is formed along the southern coast of Turkey as a consequence of the cold outbreaks along this coast, the main source regions being located near the island of Rhodes, Gulf of Antalya and to the north of Cyprus, with a source region of secondary importance near Egypt (Wüst, 1960; Morcos, 1972; Özturgut, 1976).

The various oceanographic cruises in the past have covered the eastern Mediterranean in a sparse manner, yielding only descriptive information on the spreading of the LIW core layer. For example, only 48 winter stations were available to Wüst (1961). The only outstanding experiment designed to locate regions of LIW formation and to determine distributions at the source is due to Özturgut (1976). This cruise was intentionally planned in the months of February and March (a period missed by previous cruises due to weather conditions) since these months are the sinking periods of LIW (Wüst, 1961; Morcos, 1972). Özturgut's data, having great detail and precision, proved the importance of meso-scale processes. LIW sources of 10-50 km. in radii were found in the Rhodes basin and the Gulf of Antalya, extending to depths of more than 300 m, and rapid changes in the existing patterns were observed within a weeks' time.

Other details provided by Özturgut (1976) include the presence of eddies, sharp fronts and oscillatory patterns (possible internal waves) in the region, which should all be important in the energetics of LIW formation. A cyclonic cold-core eddy in the Rhodes basin (Moskalenko and Ovchinnikov, 1965; Accerboni and Grancini, 1972; Özturgut, 1976; Phillippe and Harang, 1982) and frontal zones (Levine and White, 1972; Özturgut, 1976) appear to be permanent features.

Although these LIW source regions to the west of Cyprus are better known and confirmed by oceanographic data, the coverage for the eastern part of the NE Mediterranean (Cilician basin and Gulf of İskenderun) is generally insufficient, although strong reasons exist for expecting source regions in this area (section 2, Özsoy et al., 1980). For example, Moskalenko and Ovchinnikov (1965) present evidence of LIW spreading from the area to the north of Cyprus, but these findings are also far from being conclusive in the absence of detailed coverage in this area.

Northerly outbreaks of cold, dry continental air masses in southern Aegean sea and along the three mountain gaps in southern Asia Minor (section 2) create favorable conditions for LIW formation in the NE Mediterranean. A comparison of the regimes of Gulf of Antalya, Göksu valley and Gulf of İskenderun shows that Poyraz winds at Göksu valley could be the most effective among these wind regimes (Özsoy et al, 1980). As a result of the efficient heat, mass and momentum transfer associated with Poyraz winds (section 2, Fig 4) the highest rate of evaporation from the sea surface and deep mixing is expected near the Göksu valley followed next by the Antalya region. Near Iskenderun and Antalya the net evaporation is much reduced due to orographic rainfall (Özsoy et al, 1980).

Annual variation of surface properties observed at the southern coast of Asia Minor to the northeast of Cyprus (Özsoy et al, 1980) indicate that the highest densities are reached during the months of February-March, while the surface temperature reaches its minimum value. Surface salinity reaches a maximum in February, then during the following months of April and May falls to its minimum. In February the temperature, salinity and density values are close to the values with which LIW is identified ($T = 15.5^{o}C$, $S = 39.1$ %0, $\sigma_t = 29.0$, Morcos, 1972). During the following month, the surface dense waters sink and are replaced by the minimum salinity (38.6 %0) Atlantic water underlying the seasonal thermocline (at a depth of 60-100 m) and above the maximum salinity LIW (at a depth of 250-300 m), (Morcos, 1972; Lacombe and Tchernia, 1974).

8. RECOMMENDATIONS

The atmospheric and oceanographic observations in the northeastern Mediterranean display energetic motions within various space-time scales. Experimental designs with recognition of these scales should be of central importance for a better understanding of the dynamical aspects, while the need for more detailed observations and models of the various processes can not be overemphasized. An aspect, believed to be at least partially conveyed by this review, is the key role of meso-scale processes in the region, with possible extension to the Mediterranean oceanography in general.

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TABLE 1.a

CURRENTS AT ERDEMLI

(Site E)

Measurement Period (Month/Day/Year)	Data Length (Hours)	Depth (m)	Mean Current (cm/s) $\bar{u}_R$	$\bar{v}_R$	Standard Deviation (cm/s) σ_{u_R}	σ_{v_R}
3/22/78 - 4/6 /78	363	5	- 8.3	0.3	12.9	5.6
3/22/78 - 4/6 /78	363	10	- 4.6	0.2	10.8	5.5
8/26/78 - 9/29/78	822	10	- 9.4	-1.2	13.4	7.1
10/4 /78 - 10/30/78	628	15	0.9	1.2	16.0	2.9
11/7 /78 - 12/1 /78	590	15	4.4 } (*)	1.7 } (*)	24.8	4.0
12/9 /78 - 12/29/78	504	15	-12.7	0.6	15.5	4.1
12/31/78 - 1 /31/79	768	15	- 4.4	0.6	15.9	3.2
6 /6 /80 - 8 /19/80	1777	20	-13.7	0.1	9.4	3.1
		Overall mean :	- 7.5	0.3	14.6	4.4

(*) Means for 5 months

$\bar{u}_R = -4.3$, $\bar{v}_R = 0.5$

TABLE 1.b

CURRENTS AT AKKUYU

(Site A)

Meassurement Period (Month/Day/Year)	Data Length (hours)	Depth (m)	Mean current (cm/s) $\bar{u}$	$\bar{v}$	Standard Deviation (cm/s) σ_u	σ_v
8/23/77 - 9/11/77	490	5	0.4	-0.1	6.9	5.6
8/23/77 - 10/14/77	1231	15	-0.2	0.2	7.8	6.9
10/2 /77 - 10/23/77	513	5	1.1	1.0	6.1	5.9
10/24/77 - 12/12/77	1153	15	-0.5	-0.2	4.3	2.6
2/13/78 - 3/14/78	746	5	3.3	2.1	5.9	4.8
2/13/78 - 3/14/78	746	15	0.3	1.9	6.8	5.0
4/9 /78 - 5/13/78	856	5	0.3	-4.1	17.4	13.1
4/9 /78 - 5/13/78	826	5	-0.6	-1.0	11.3	9.5
5/18/78 - 6/19/78	806	5	-2.7	-0.9	8.5	5.8
5/18/78 - 6/19/78	860	15	-1.3	-0.8	9.7	6.1
6/19/80 - 9/4 /80	1878	20	-0.7	2.2	7.7	8.0
6/19/80 - 9/20/80	2269	20	-2.0	3.5	9.1	7.4
		Overall mean :	-0.5	0.8	9.0	7.3

FIGURE LEGENDS

Figure 1. Tracks of Mediterranean depressions with annual frequencies (After Hydrographer of the Navy, 1976).

Figure 2. Cross sections of land topography 100 miles inland from the coast along (a) the northern and (b) the southern shores of the Mediterranean sea (After Reiter, 1975).

Figure 3. Land topography above 1000 m. and the channelization of northerly winds. (After Özsoy, 1981).

Figure 4. Time-series of wind stress, relative humidity and sea water temperature (After Özsoy, 1981).

Figure 5. Monthly mean wind stress at four coastal meteorology stations in the Cilcian basin (After Ataktürk, 1980; Özsoy, 1981).

Figure 6.a. Wind stress rotary spectra for four coastal stations in August (1977), spectral density units $6x10^4 m^4 s^{-4} cph^{-1}$ (After Ataktürk, 1980).

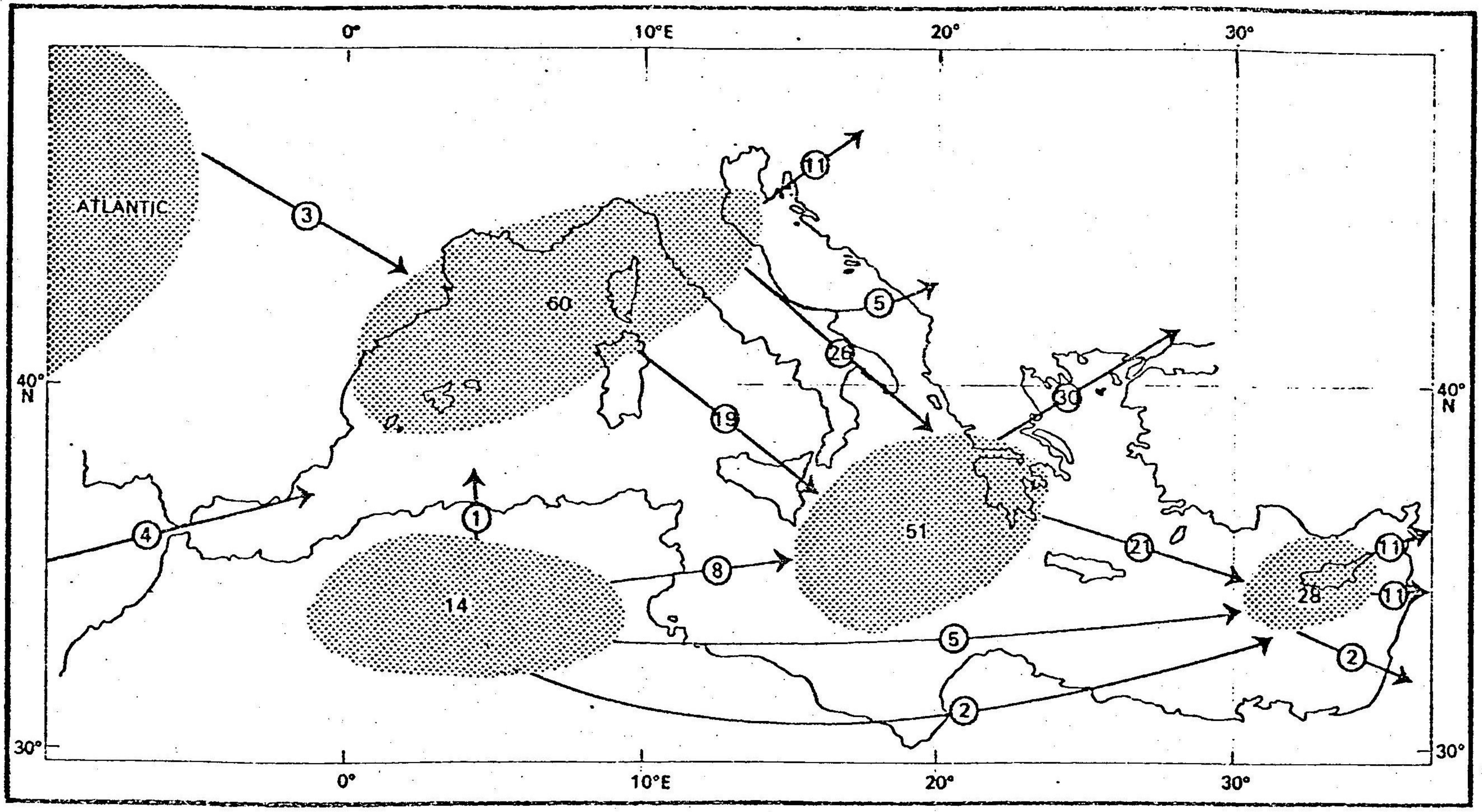
Figure 6.b. Wind stress rotary spectra for four coastal stations in November (1977), spectral density units $6x10^4 m^4 s^{-4}$ cph^{-1} (After Ataktürk, 1980).

Figure 7. Bottom topography in the eastern Mediterranean (After Stanley, 1972).

Figure 8. Mean circulation in the northeastern Mediterranean (After Collins and Banner, 1979).

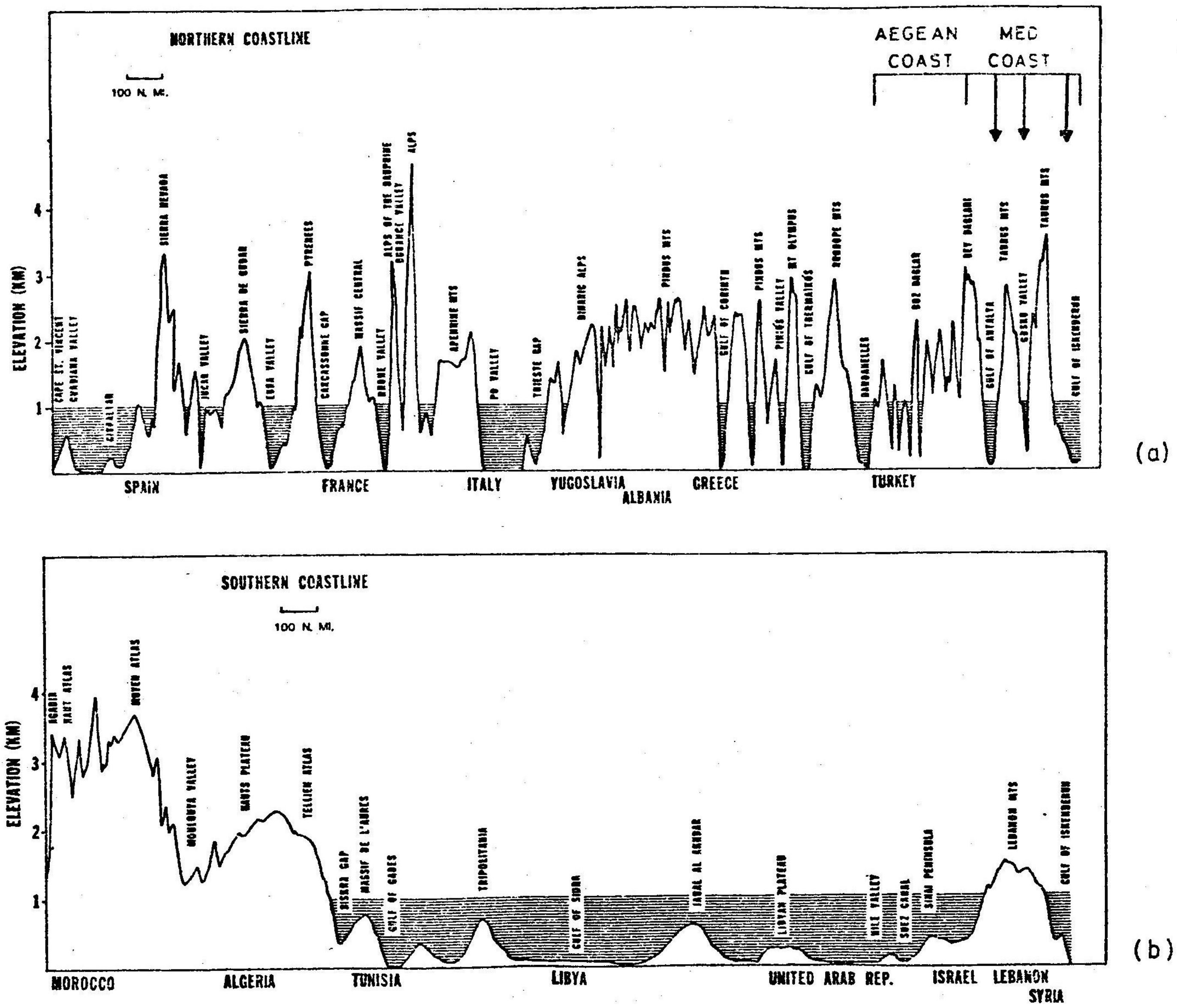
Figure 9. Current meter sites and nearshore topography (After Ünlüata, et al,1983).

Figure 10. Numerical solution for mean circulation on the continental shelf extending from Erdemli to Anamur (After Ünlüata,el al, 1983)

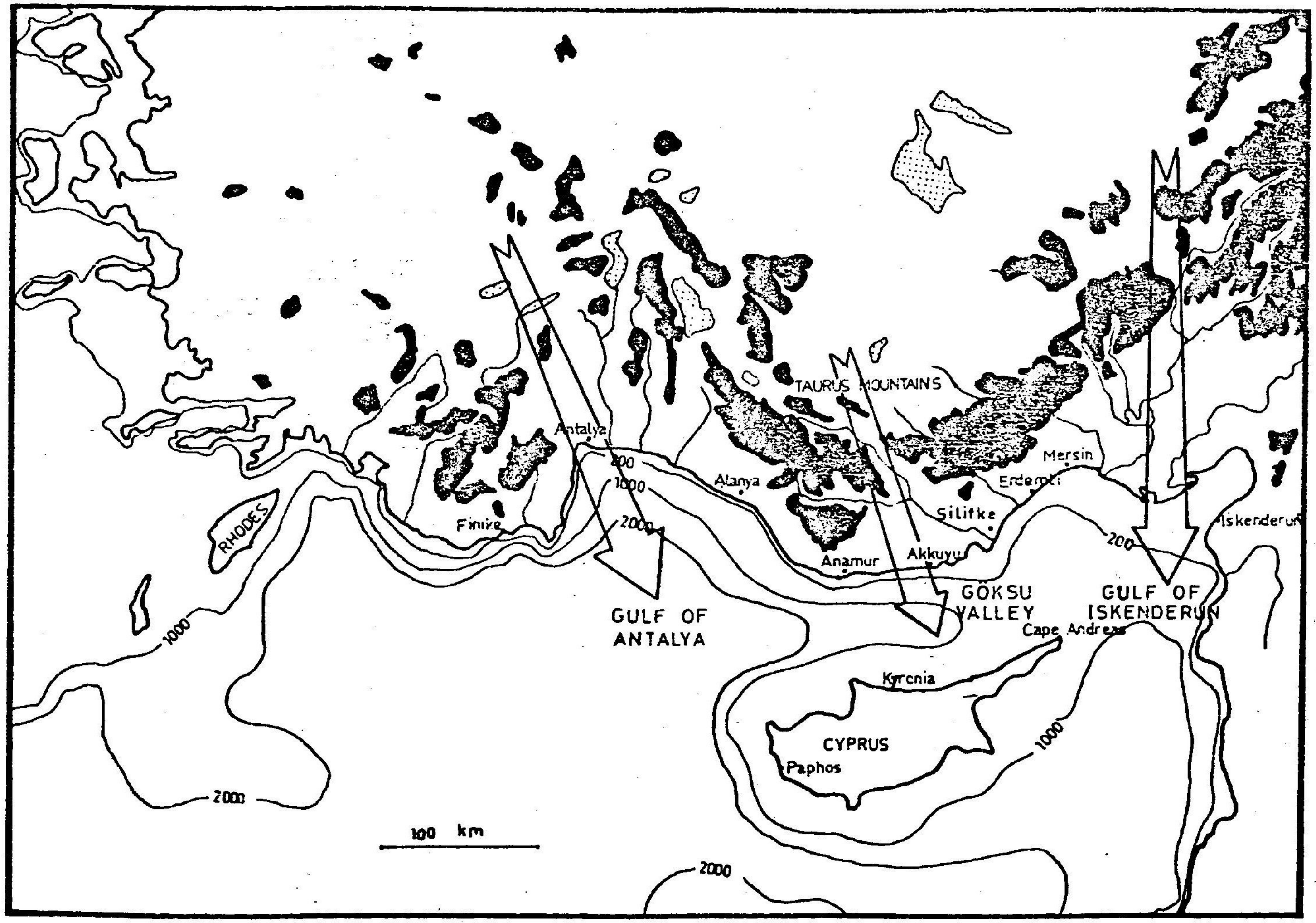


0°
10°E
20°
30°
ATLANTIC
3
60
11
5
26
19
30
40° N
40° N
4
1
8
14
51
21
28
11
11
5
2
2
30°
30°
0°
10°E
20°
30°

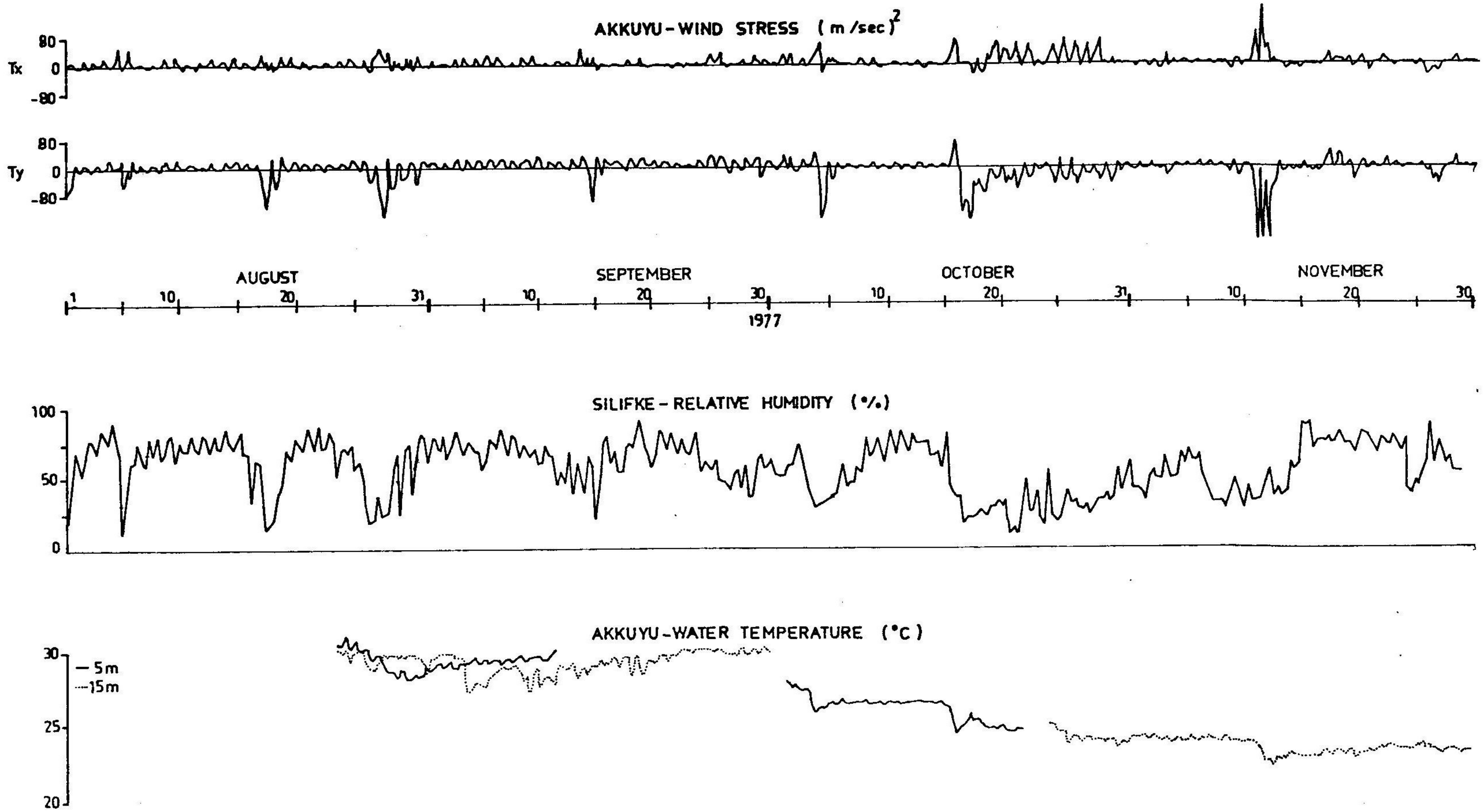
1



NORTHERN COASTLINE
100 N. MI.
AEGEAN COAST
MED COAST
ELEVATION (KM)
SPAIN
FRANCE
ITALY
YUGOSLAVIA
ALBANIA
GREECE
TURKEY
(a)
SOUTHERN COASTLINE
100 N. MI.
ELEVATION (KM)
MOROCCO
ALGERIA
TUNISIA
LIBYA
UNITED ARAB REP.
ISRAEL
LEBANON
SYRIA
(b)

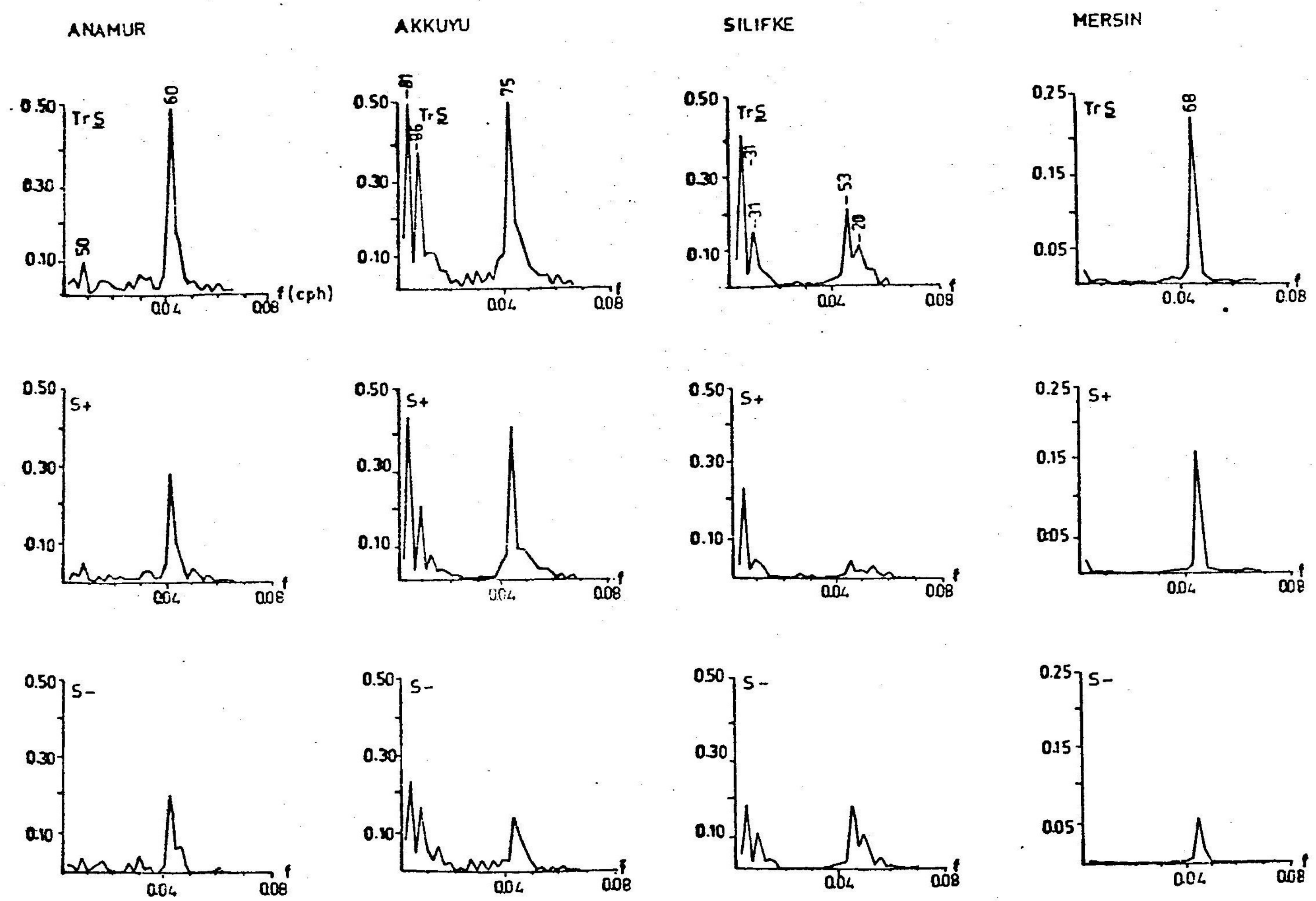

TAURUS MOUNTAINS
Antalya
Alanya
Mersin
Erdemli
Silifke
Anamur
Akkuyu
Finike
Iskenderun
RHODES
GULF OF
ANTALYA
GÖKSU
VALLEY
GULF OF
ISKENDERUN
Cape Andreas
Kyrenia
CYPRUS
Paphos
200
1000
2000
100 km

3

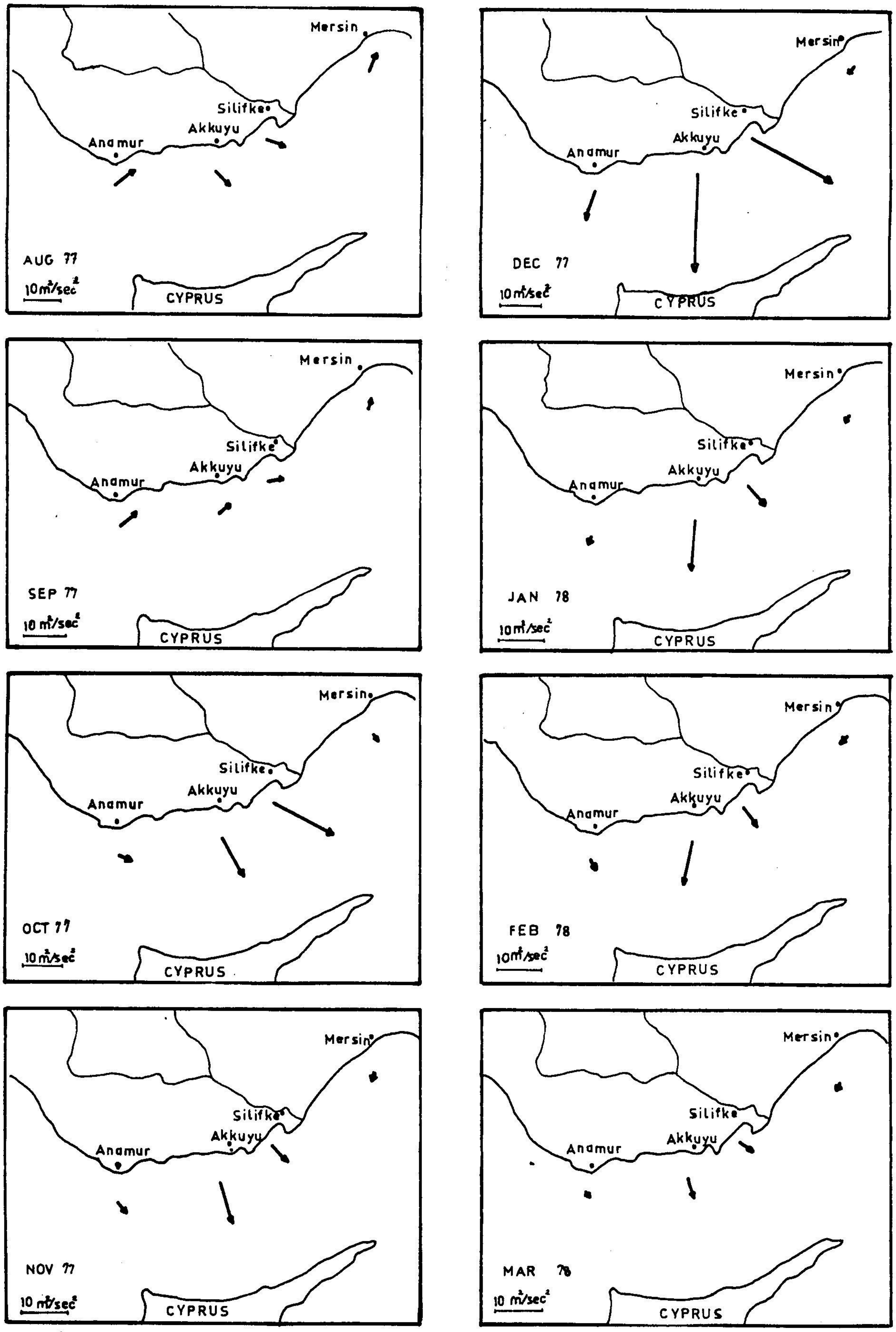

AKKUYU-WIND STRESS (m/sec)²
Tx
Ty
80
0
-80
AUGUST
SEPTEMBER
OCTOBER
NOVEMBER
1
10
20
31
30
1977
SILIFKE-RELATIVE HUMIDITY (%)
100
50
0
AKKUYU-WATER TEMPERATURE (°C)
30
25
20
— 5m
---15m

4

AUGUST 1977

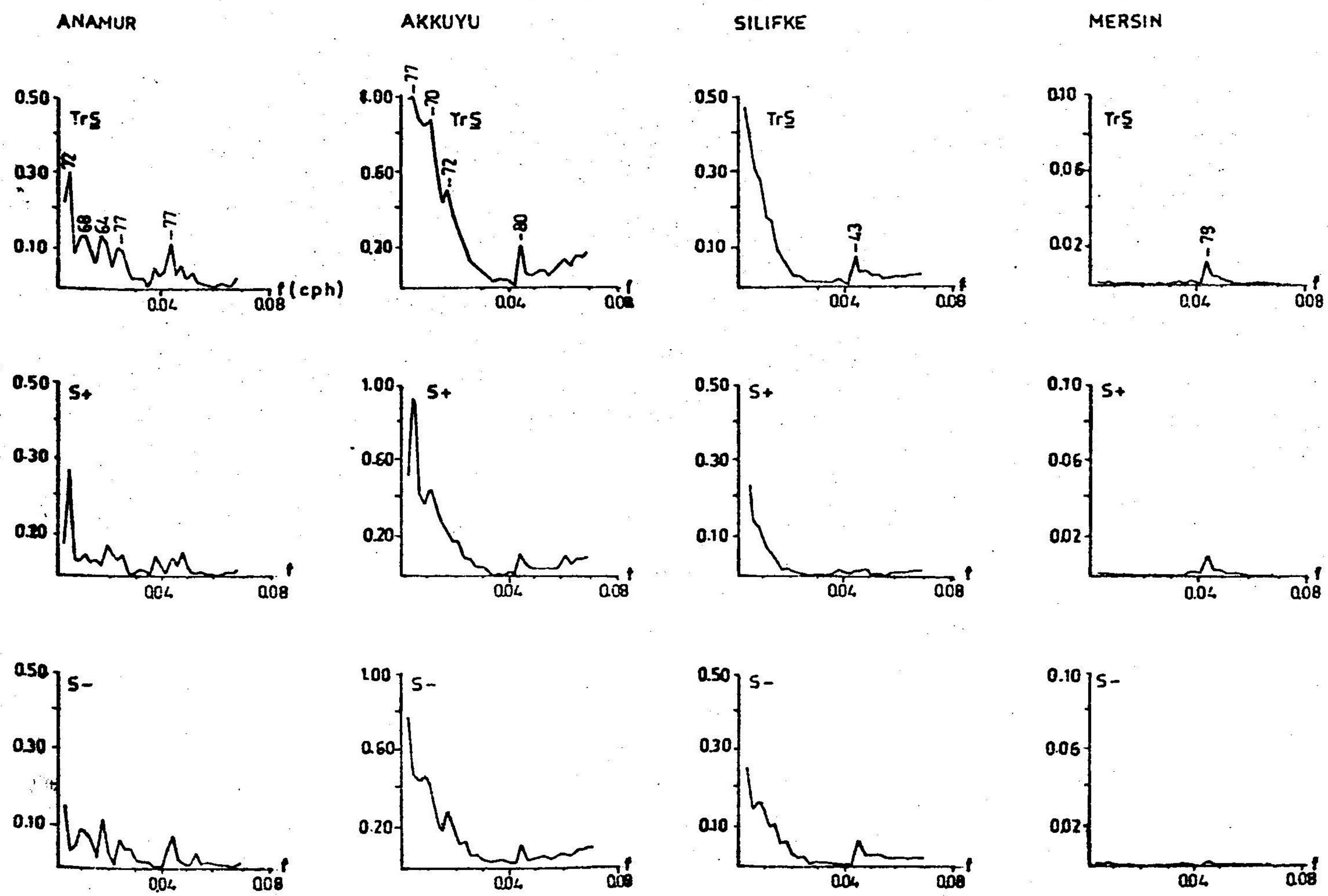


(a)

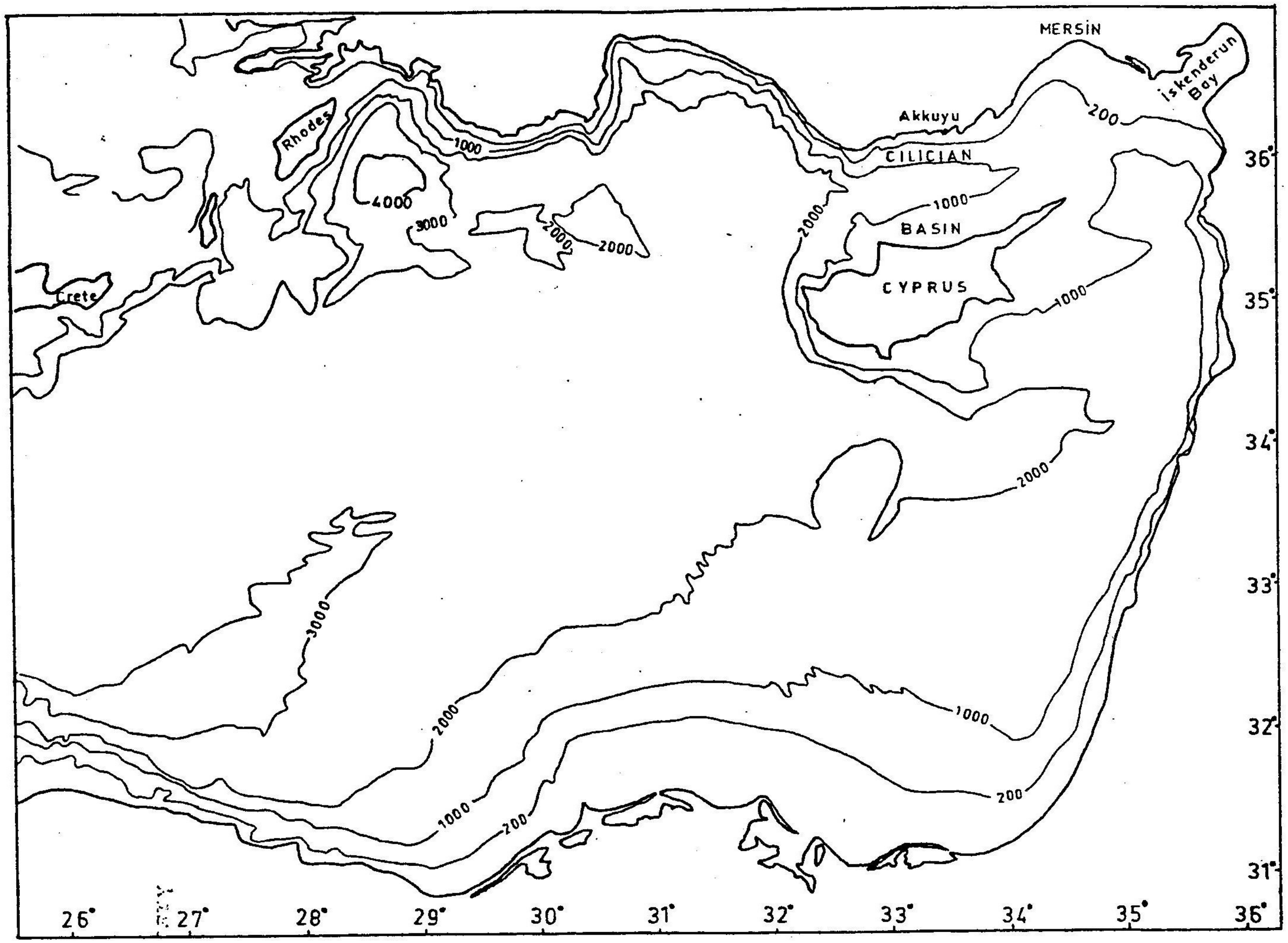


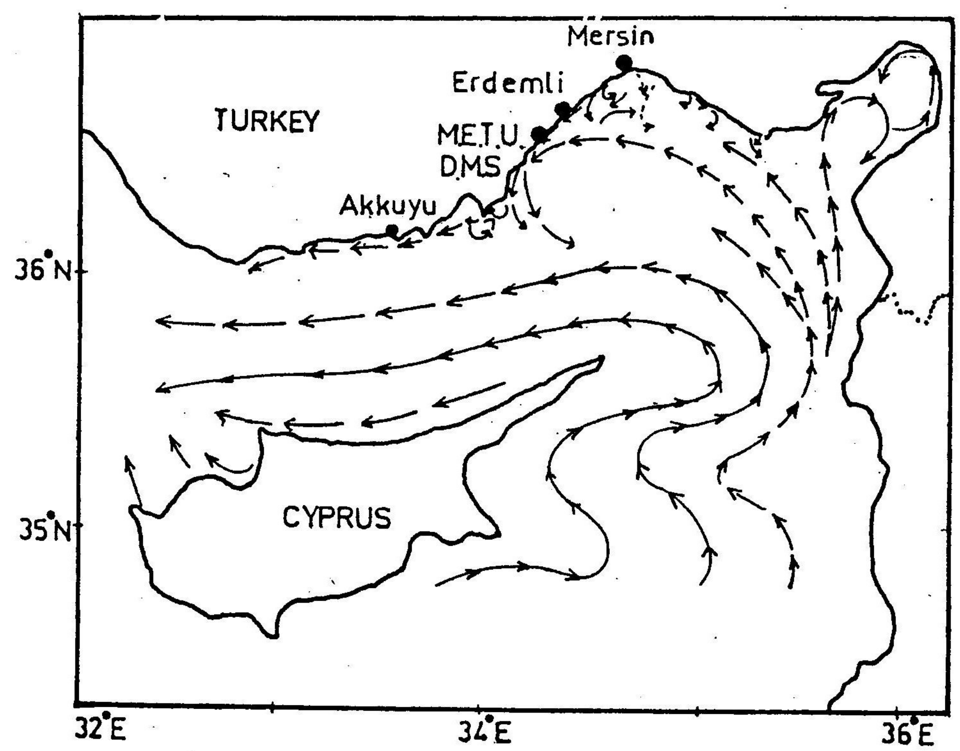
Mersin
Silifke
Akkuyu
Anamur
AUG 77
10m²/sec²
CYPRUS
Mersin
Silifke
Akkuyu
Anamur
DEC 77
10m²/sec²
CYPRUS
Mersin
Silifke
Akkuyu
Anamur
SEP 77
10 m²/sec²
CYPRUS
Mersin
Silifke
Akkuyu
Anamur
JAN 78
10m²/sec²
CYPRUS
Mersin
Silifke
Akkuyu
Anamur
OCT 77
10 m²/sec²
CYPRUS
Mersin
Silifke
Akkuyu
Anamur
FEB 78
10m²/sec²
CYPRUS
Mersin
Silifke
Akkuyu
Anamur
NOV 77
10 m²/sec²
CYPRUS
Mersin
Silifke
Akkuyu
Anamur
MAR 78
10 m²/sec²
CYPRUS

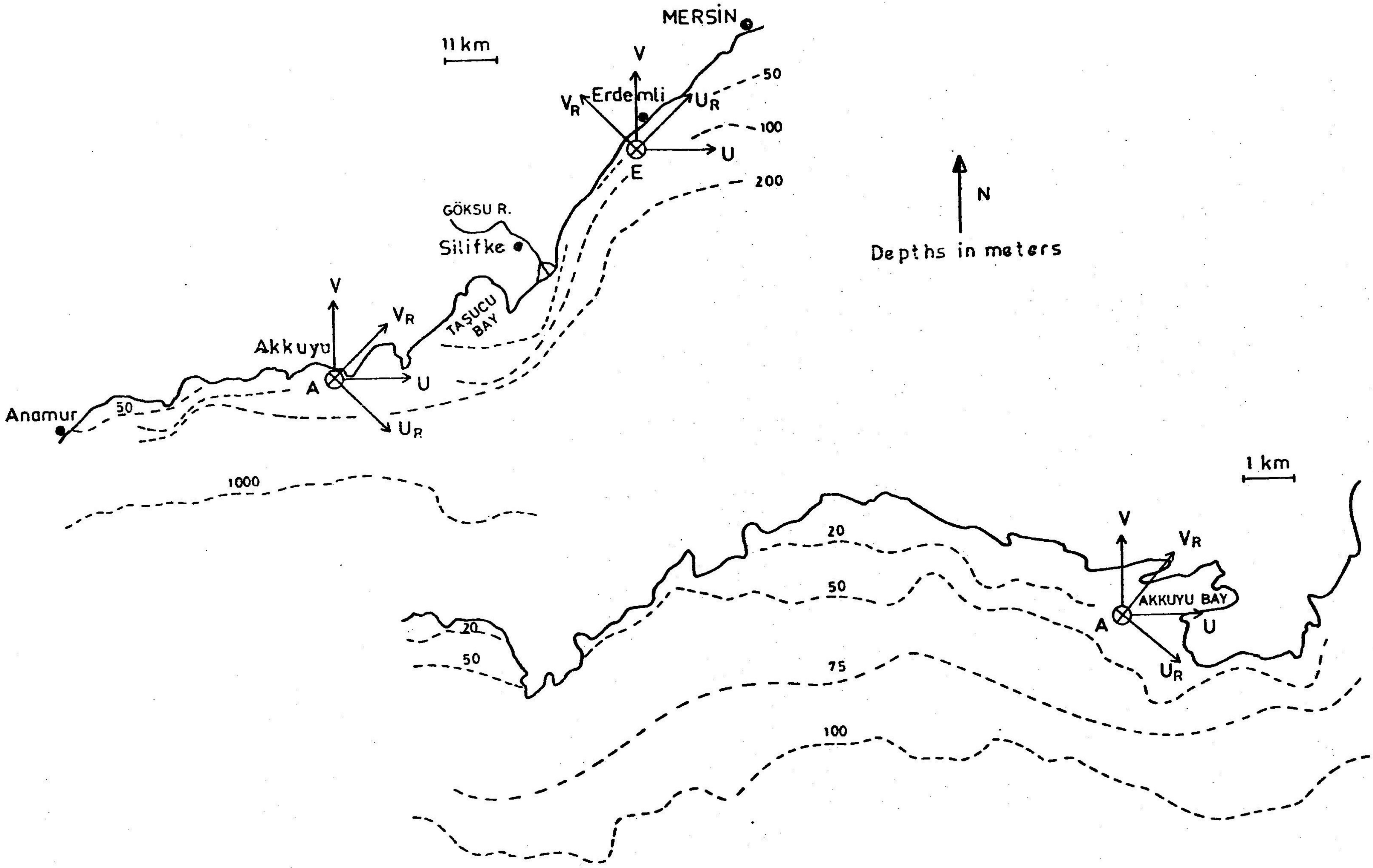
NOVEMBER 1977



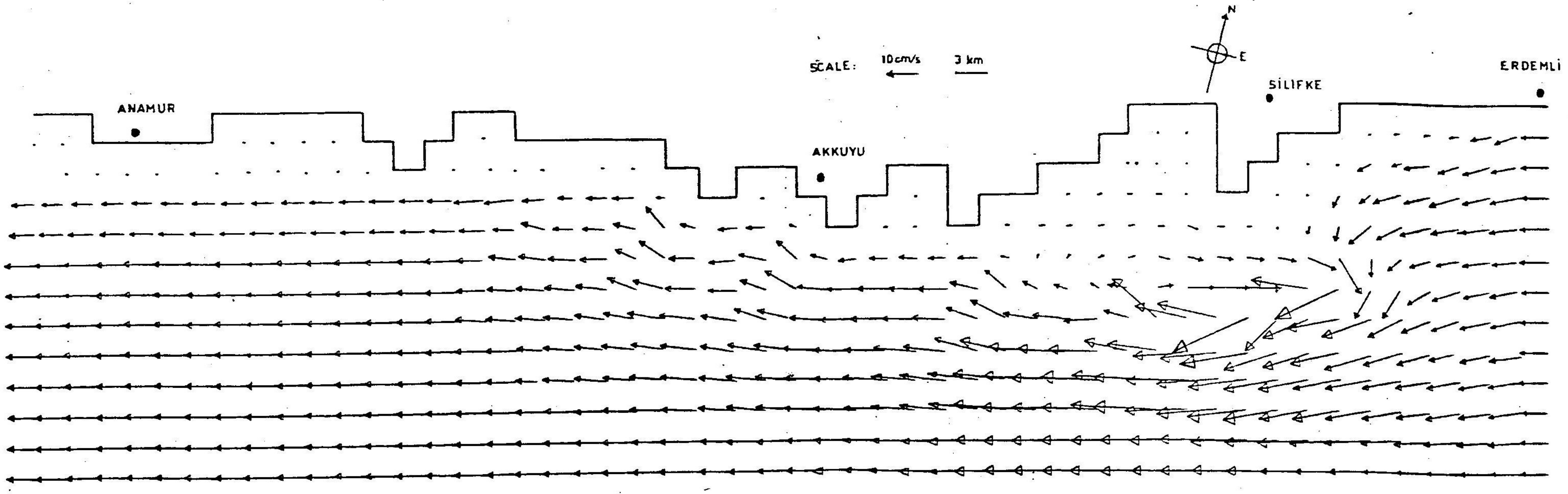
(d)

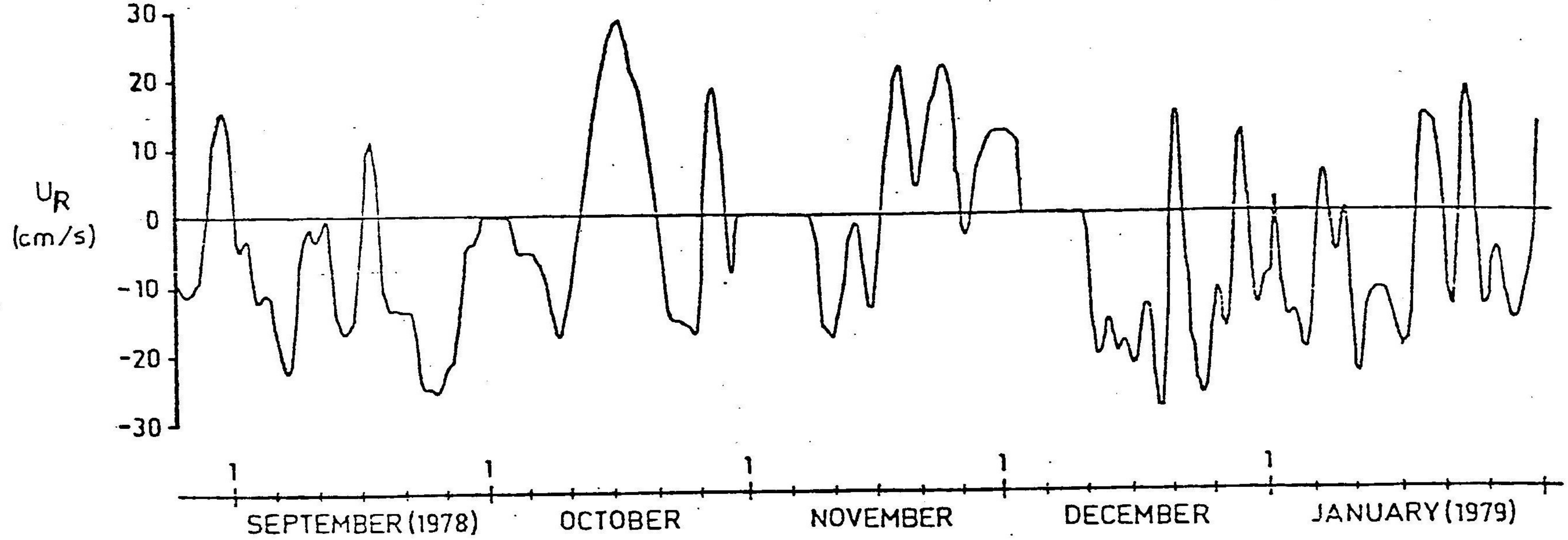

MERSİN
İskenderun Bay
Akkuyu
200
1000
Rhodes
CILICIAN
1000
BASIN
4000
3000
2000
2000
2000
Crete
CYPRUS
1000
36°
35°
34°
2000
33°
3000
2000
1000
32°
200
1000
200
31°
26°
27°
28°
29°
30°
31°
32°
33°
34°
35°
36°


Mersin
Erdemli
TURKEY
M.E.T.U.
D.M.S.
Akkuyu
36°N
35°N
CYPRUS
32°E
34°E
36°E

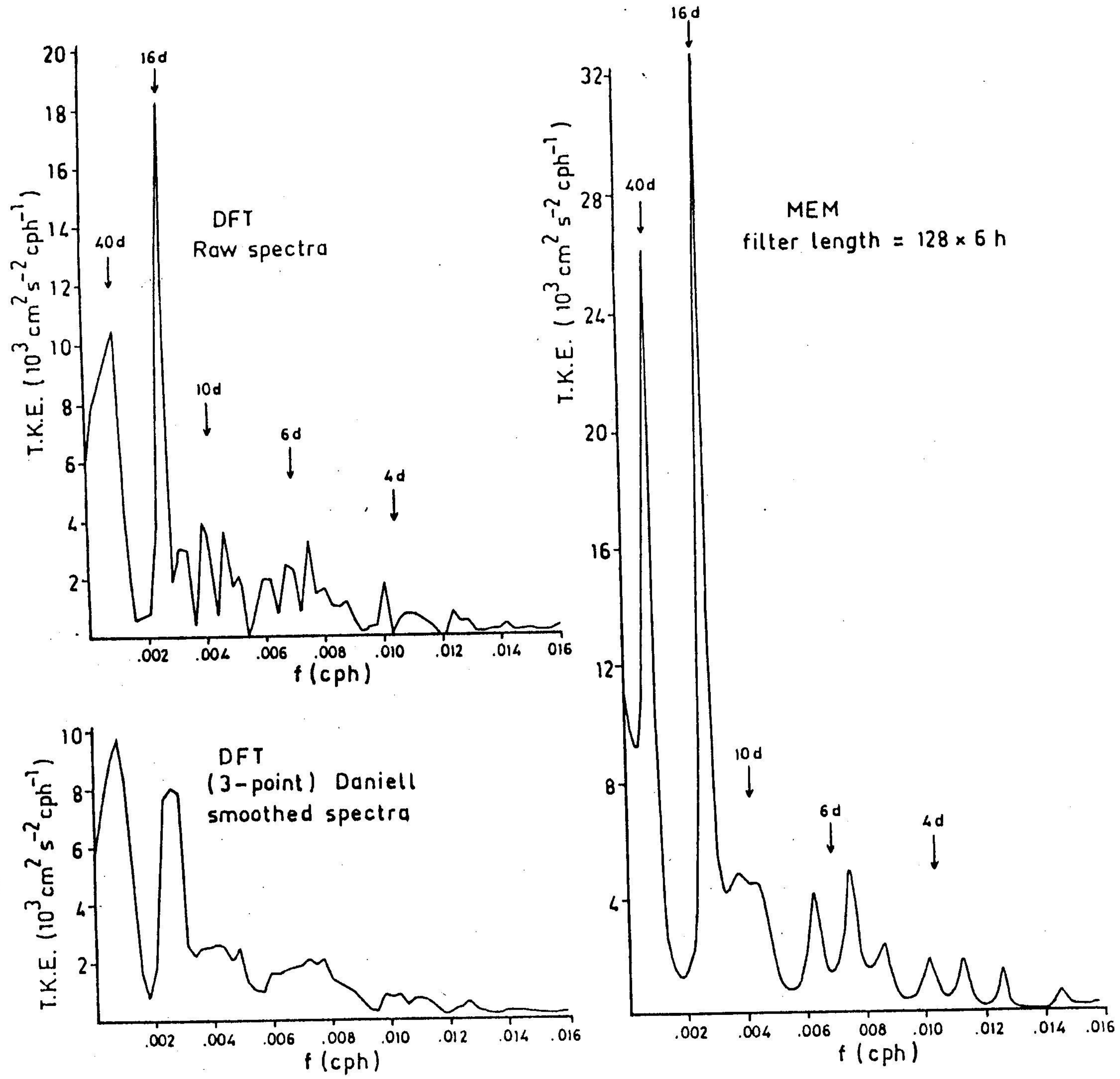


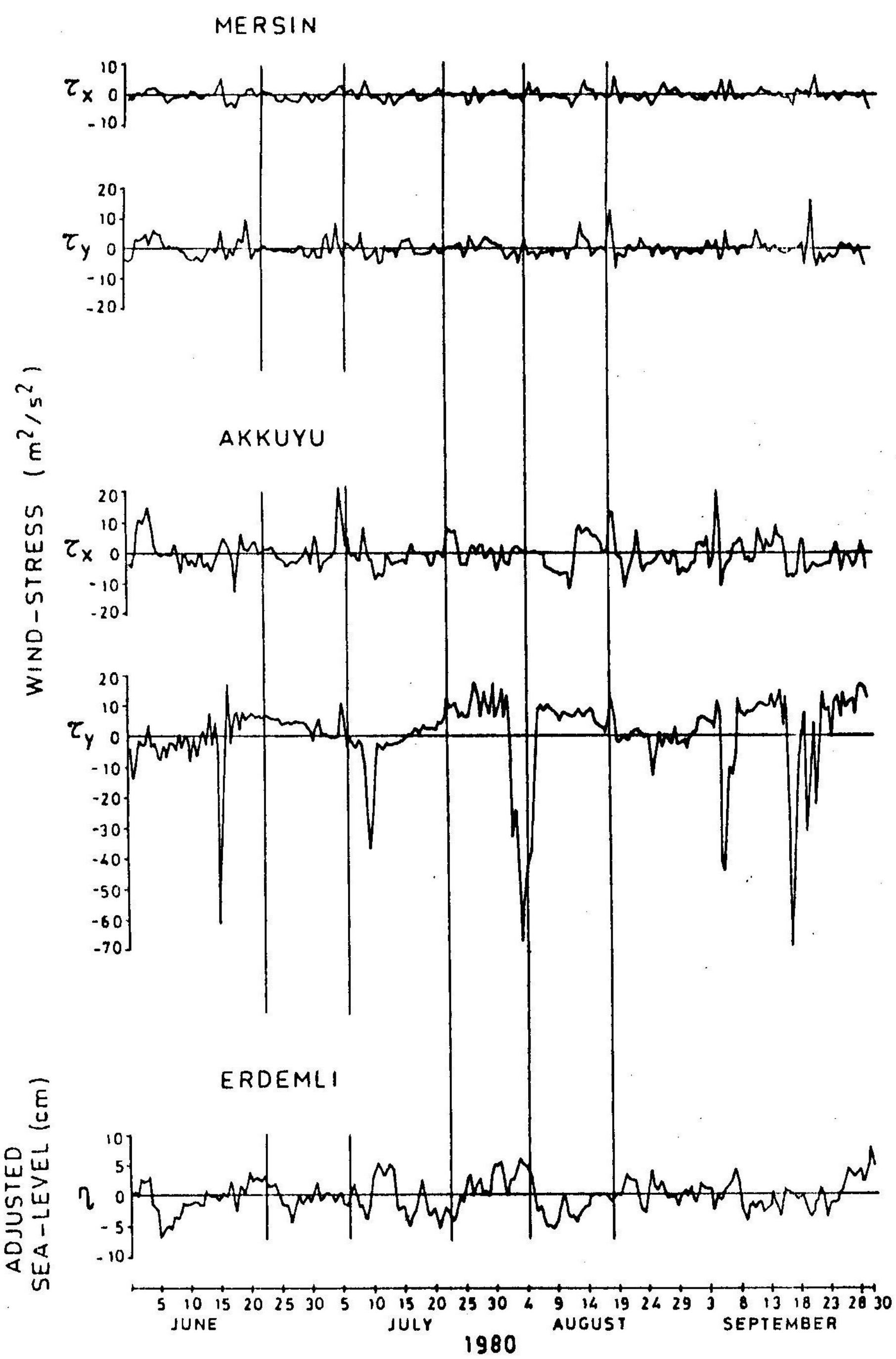
MERSIN
11 km
V
V
R
Erdemli
U
R
50
100
U
E
200
N
Depths in meters
GÖKSU R.
Silifke
V
V
R
TAŞUCU BAY
Akkuyu
A
U
Anamur
50
U
R
1000
1 km
V
V
R
20
50
AKKUYU BAY
A
U
20
50
75
U
R
100


SCALE:
10cm/s
3 km
N
E
ANAMUR
AKKUYU
SİLİFKE
ERDEMLİ

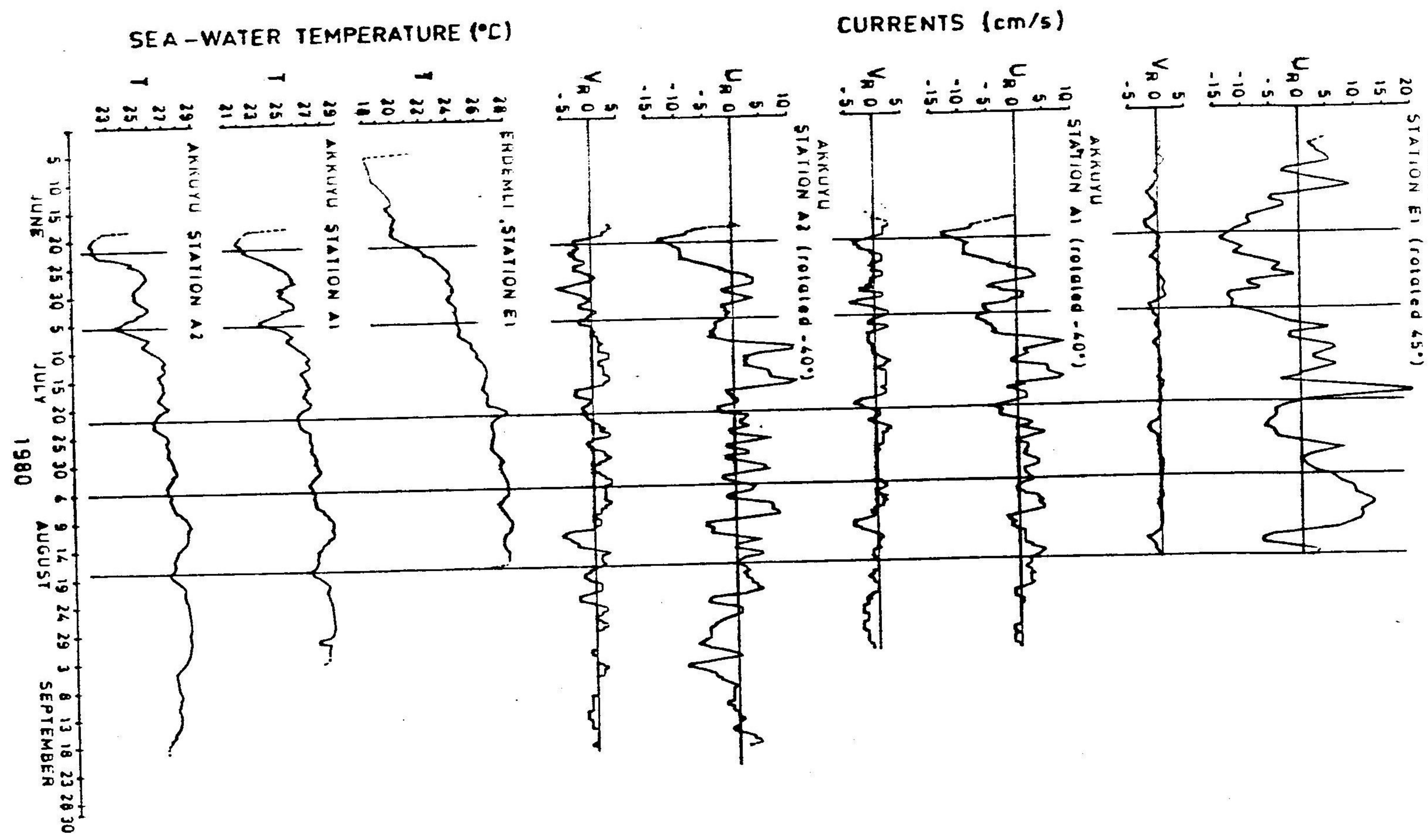


30
20
10
0
-10
-20
-30
U_R
(cm/s)
1
1
1
1
1
SEPTEMBER (1978)
OCTOBER
NOVEMBER
DECEMBER
JANUARY (1979)

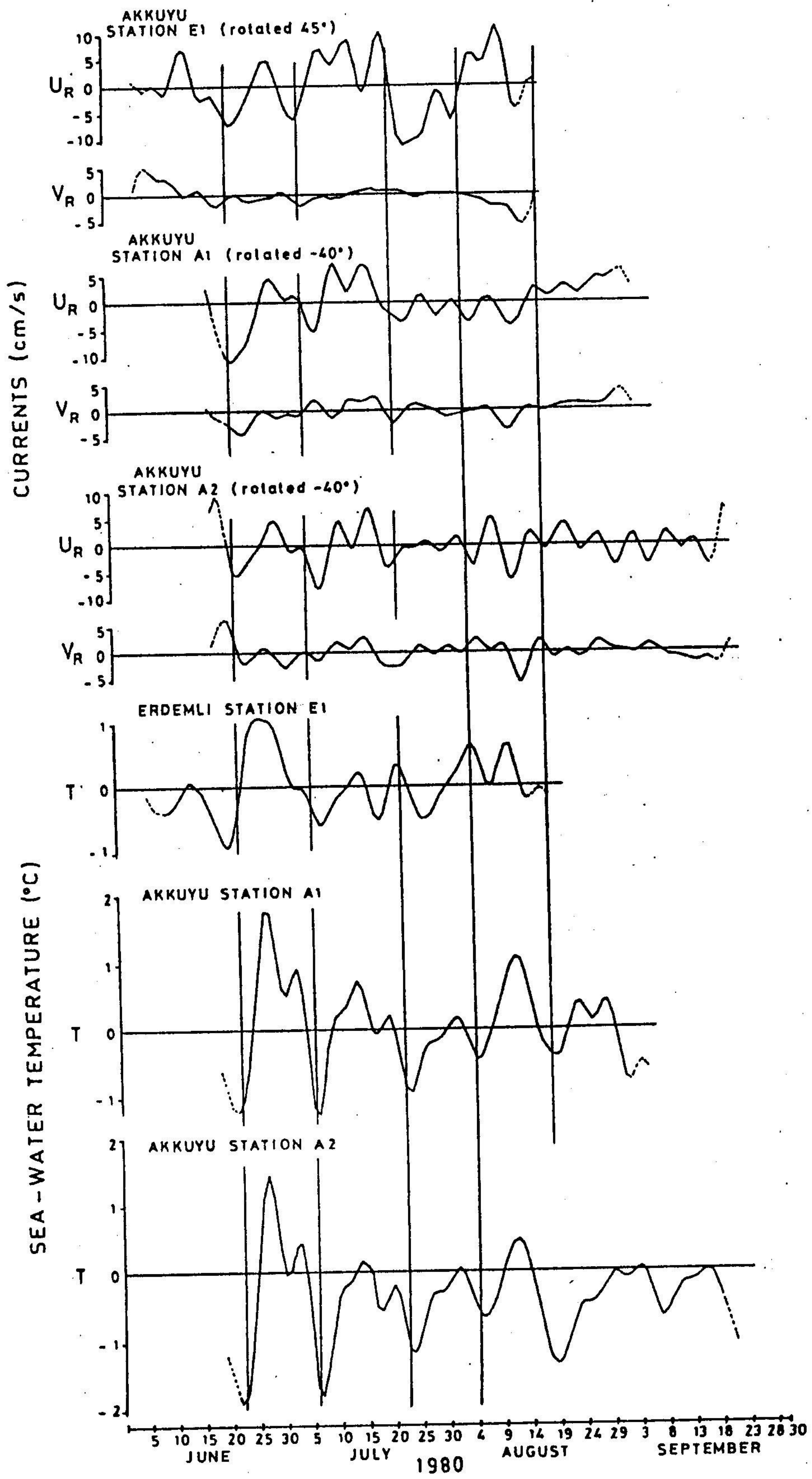

DFT
Raw spectra
40d
16d
10d
6d
4d
T.K.E. ($10^3 cm^2 s^{-2} cph^{-1}$)
f (cph)
.002 .004 .006 .008 .010 .012 .014 .016
DFT
(3-point) Daniell
smoothed spectra
T.K.E. ($10^3 cm^2 s^{-2} cph^{-1}$)
f (cph)
MEM
filter length = 128 × 6 h
40d
16d
10d
6d
4d
T.K.E. ($10^3 cm^2 s^{-2} cph^{-1}$)
f (cph)



MERSIN
AKKUYU
ERDEMLI
WIND-STRESS (m²/s²)
ADJUSTED SEA-LEVEL (cm)
JUNE
JULY
AUGUST
SEPTEMBER
1980

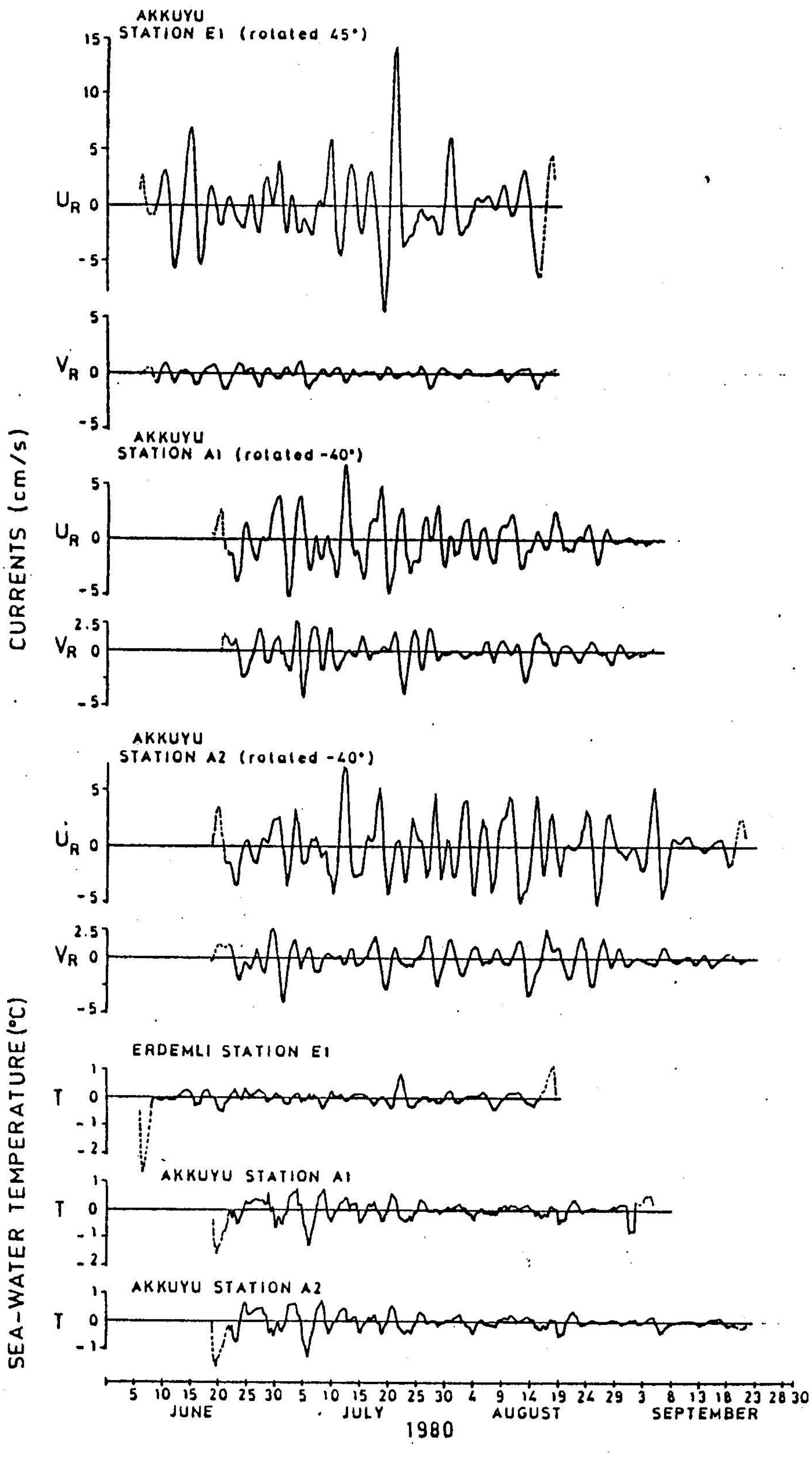


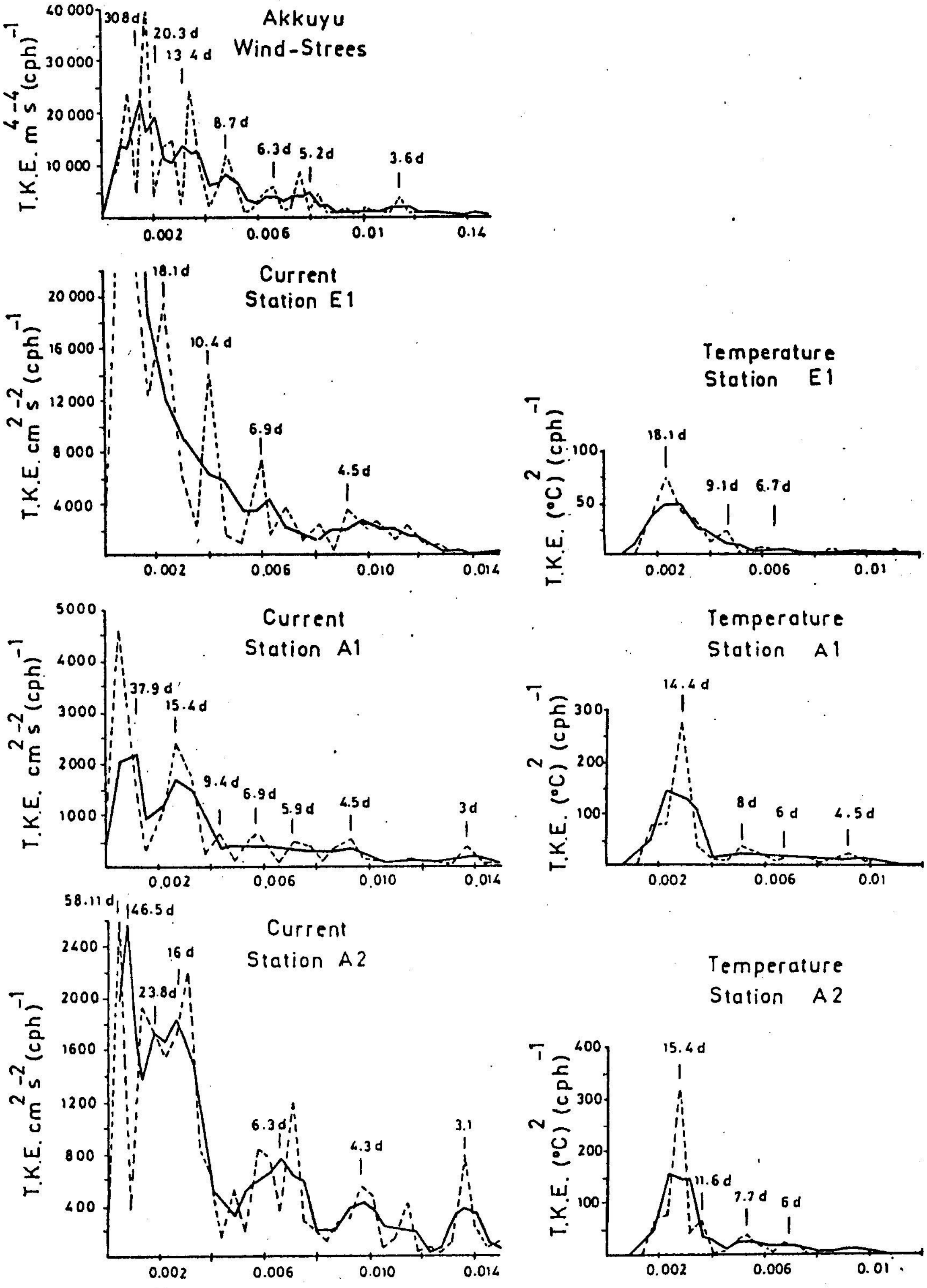
CURRENTS (cm/s)
SEA-WATER TEMPERATURE (°C)
AKKUYU STATION E1 (rotated 45°)
AKKUYU STATION A1 (rotated -40°)
AKKUYU STATION A2 (rotated -40°)
ERDEMLI, STATION E1
AKKUYU STATION A1
AKKUYU STATION A2
JUNE
JULY
AUGUST
SEPTEMBER
1980

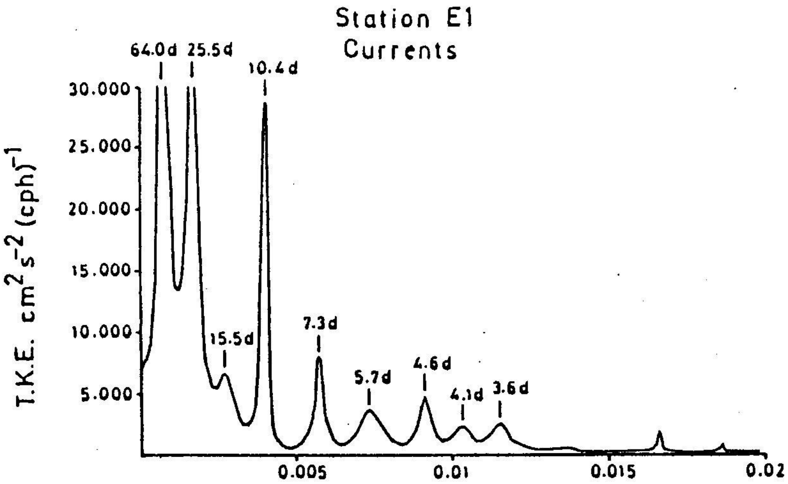


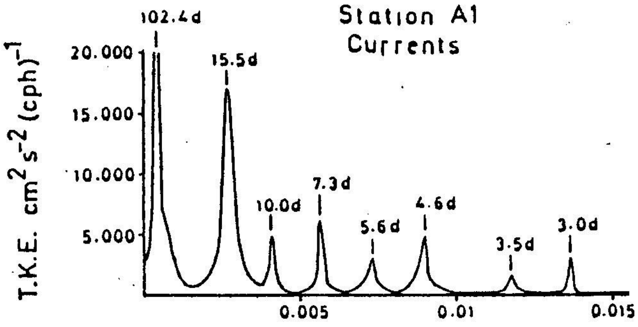
AKKUYU STATION E1 (rotated 45°)
AKKUYU STATION A1 (rotated -40°)
AKKUYU STATION A2 (rotated -40°)
CURRENTS (cm/s)
ERDEMLI STATION E1
AKKUYU STATION A1
AKKUYU STATION A2
SEA-WATER TEMPERATURE (°C)
JUNE
JULY
AUGUST
SEPTEMBER
1980

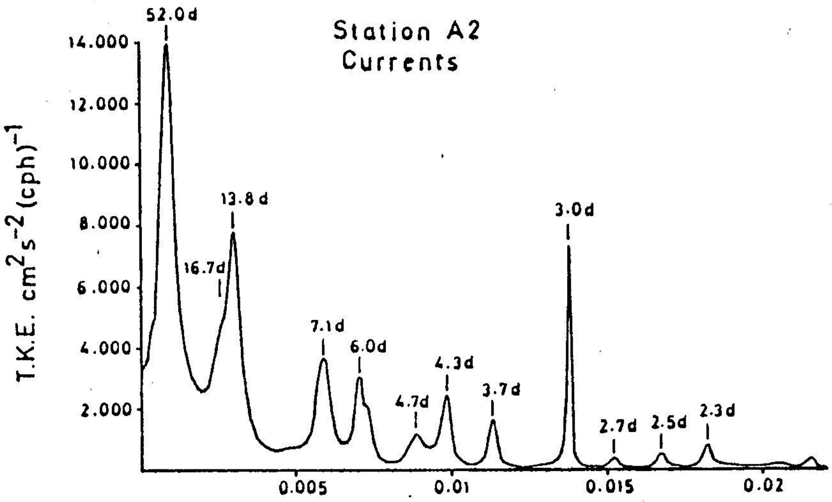
14

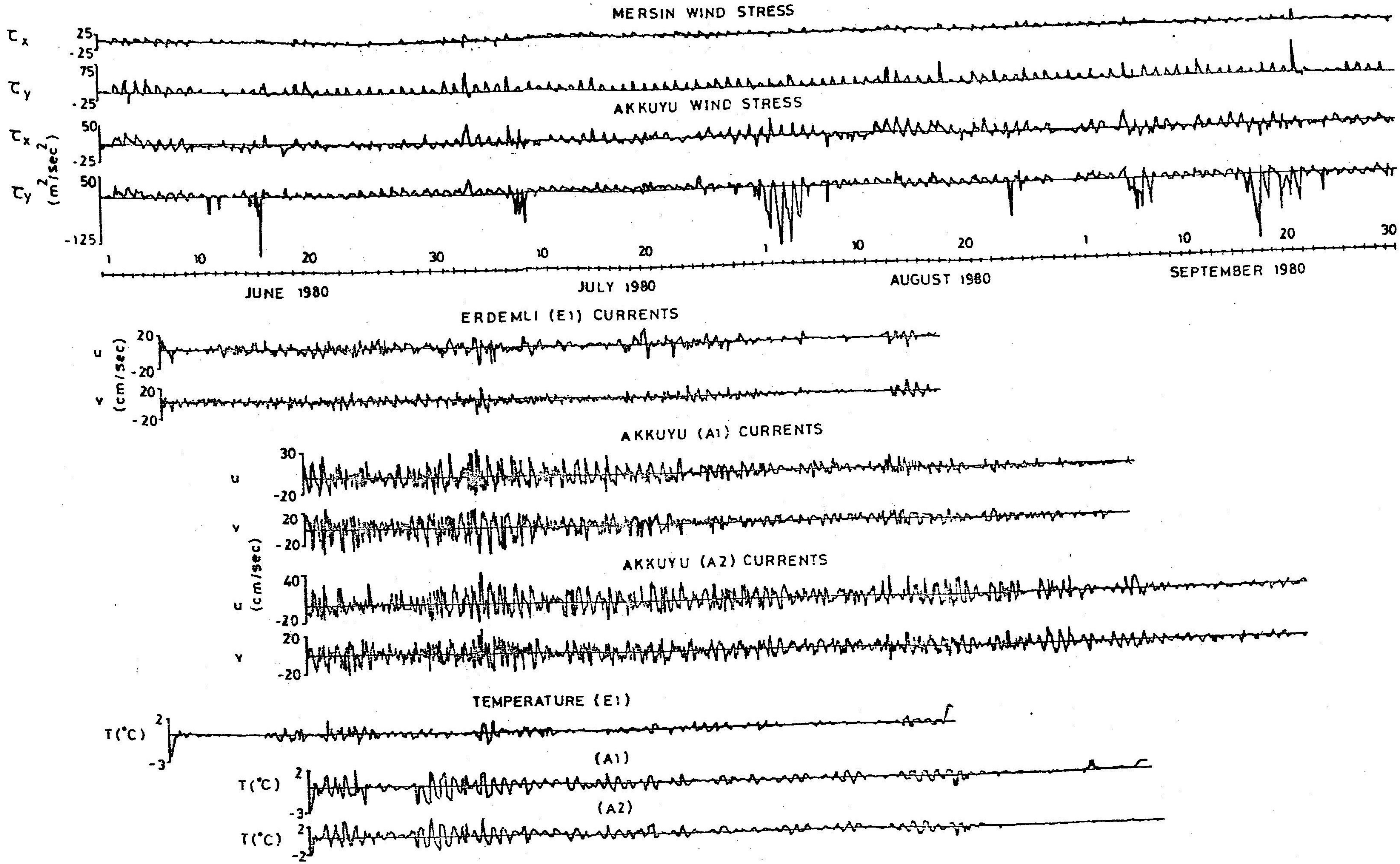

AKKUYU
STATION E1 (rotated 45°)
AKKUYU
STATION A1 (rotated -40°)
AKKUYU
STATION A2 (rotated -40°)
ERDEMLI STATION E1
AKKUYU STATION A1
AKKUYU STATION A2
CURRENTS (cm/s)
SEA-WATER TEMPERATURE (°C)
U_R
V_R
T
JUNE
JULY
AUGUST
SEPTEMBER
1980

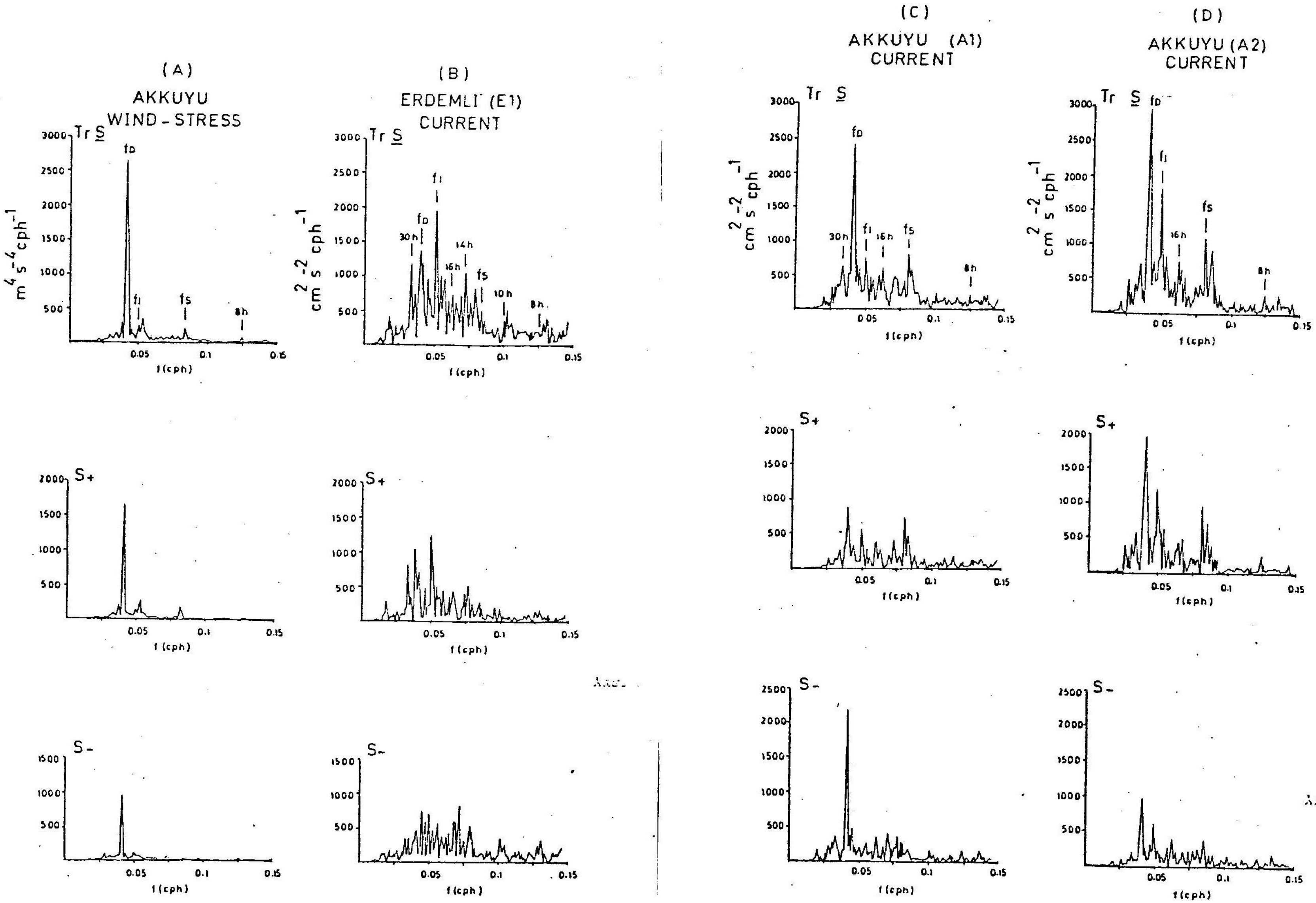

Akkuyu
Wind-Strees
T.K.E. m^4 s^-4 (cph)^-1
40 000
30 000
20 000
10 000
308d
20.3 d
13.4 d
8.7 d
6.3 d
5.2 d
3.6 d
0.002
0.006
0.01
0.14
Current
Station E1
T.K.E. cm^2 s^-2 (cph)^-1
20 000
16 000
12 000
8 000
4000
18.1 d
10.4 d
6.9 d
4.5 d
0.002
0.006
0.010
0.014
Temperature
Station E1
T.K.E. (°C)^2 (cph)^-1
100
50
18.1 d
9.1 d
6.7 d
0.002
0.006
0.01
Current
Station A1
T.K.E. cm^2 s^-2 (cph)^-1
5000
4000
3000
2000
1000
37.9 d
15.4 d
9.4 d
6.9 d
5.9 d
4.5 d
3 d
0.002
0.006
0.010
0.014
Temperature
Station A1
T.K.E. (°C)^2 (cph)^-1
300
200
100
14.4 d
8 d
6 d
4.5 d
0.002
0.006
0.01
Current
Station A2
T.K.E. cm^2 s^-2 (cph)^-1
2400
2000
1600
1200
800
400
58.11 d
46.5 d
23.8 d
16 d
6.3 d
4.3 d
3.1
0.002
0.006
0.010
0.014
Temperature
Station A2
T.K.E. (°C)^2 (cph)^-1
400
300
200
100
15.4 d
11.6 d
7.7 d
6 d
0.002
0.006
0.01

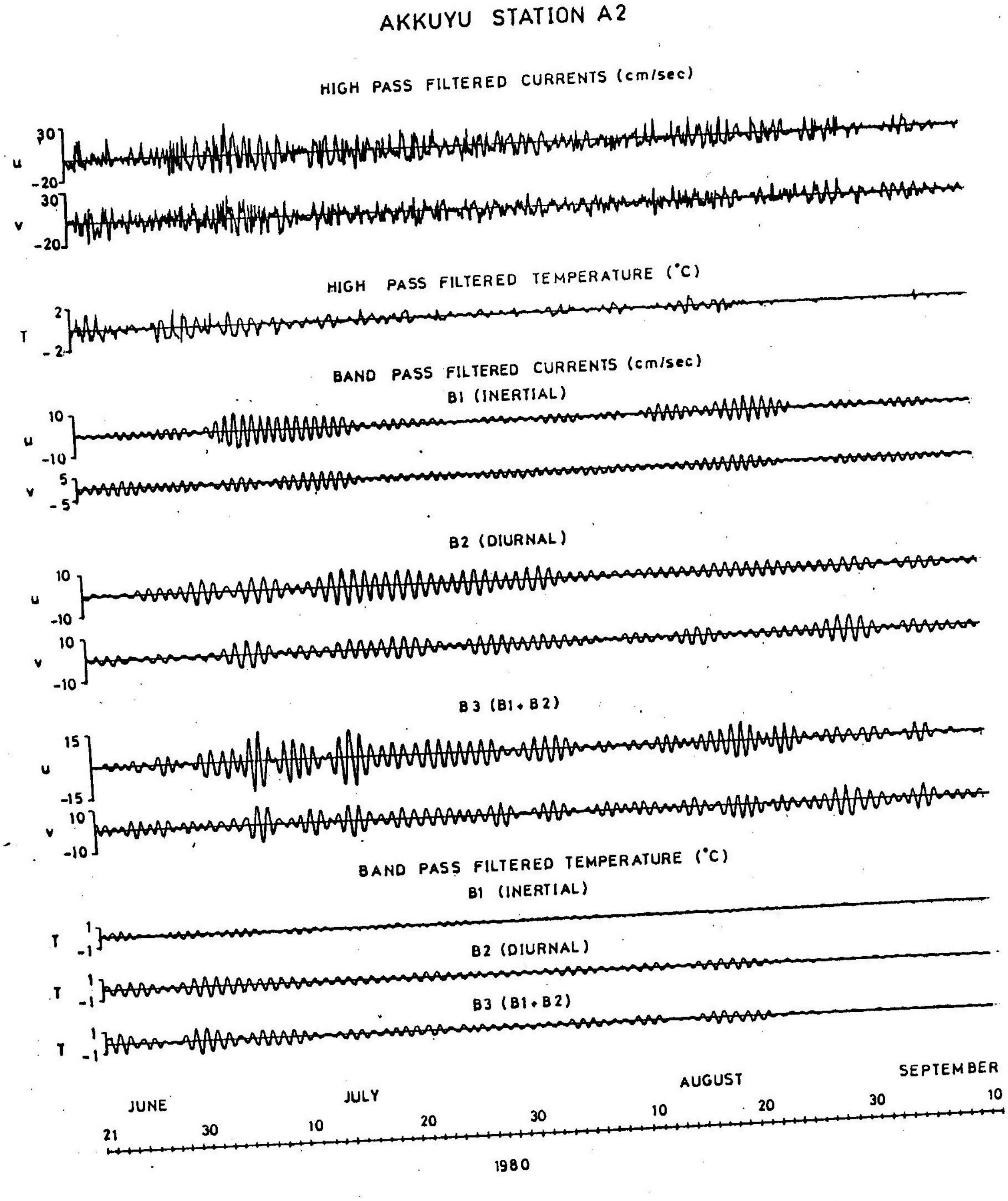

Station E1
Currents
64.0d
25.5d
10.4d
15.5d
7.3d
5.7d
4.6d
4.1d
3.6d
T.K.E. cm2s-2 (cph)-1
30.000
25.000
20.000
15.000
10.000
5.000
0.005
0.01
0.015
0.02


Station A1
Currents
102.4d
15.5d
10.0d
7.3d
5.6d
4.6d
3.5d
3.0d
T.K.E. cm2s-2 (cph)-1
20.000
15.000
10.000
5.000
0.005
0.01
0.015


Station A2
Currents
52.0d
16.7d
13.8 d
7.1d
6.0d
4.7d
4.3d
3.7d
3.0d
2.7d
2.5d
2.3d
T.K.E. cm2s-2 (cph)-1
14.000
12.000
10.000
8.000
6.000
4.000
2.000
0.005
0.01
0.015
0.02


MERSIN WIND STRESS
AKKUYU WIND STRESS
τx
τy
(m²/sec²)
25
-25
75
-25
50
-25
50
-125
1
10
20
30
JUNE 1980
JULY 1980
AUGUST 1980
SEPTEMBER 1980
ERDEMLI (E1) CURRENTS
u
v
(cm/sec)
20
-20
AKKUYU (A1) CURRENTS
30
-20
AKKUYU (A2) CURRENTS
40
TEMPERATURE (E1)
T(°C)
2
-3
(A1)
(A2)
-2


(A)
AKKUYU
WIND-STRESS
(B)
ERDEMLI (E1)
CURRENT
(C)
AKKUYU (A1)
CURRENT
(D)
AKKUYU (A2)
CURRENT
Tr S
S+
S-
fD
f1
fs
30h
16h
14h
10h
8h
f (cph)
0.05
0.1
0.15


AKKUYU STATION A2
HIGH PASS FILTERED CURRENTS (cm/sec)
u
v
30
-20
HIGH PASS FILTERED TEMPERATURE (°C)
T
2
-2
BAND PASS FILTERED CURRENTS (cm/sec)
B1 (INERTIAL)
10
-10
5
-5
B2 (DIURNAL)
B3 (B1+B2)
15
-15
BAND PASS FILTERED TEMPERATURE (°C)
B1 (INERTIAL)
B2 (DIURNAL)
B3 (B1+B2)
1
-1
JUNE
JULY
AUGUST
SEPTEMBER
21
30
10
20
1980