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## Climatic warming and accompanying changes in the ecological regime of the Black Sea during 1990s

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[1] The Black Sea ecosystem is shown to experience abrupt shifts in its all trophic levels from primary producers to apex predators in 1995–1996. It arises as a manifestation of concurrent changes in its physical climate introduced by intensive warming of its surface waters as well as abrupt increases in the mean sea level and the net annual mean fresh water flux. The warming is evident in the annual-mean sea surface temperature (SST) data by a continuous rise at a rate of  $\sim 0.25^{\circ}\text{C}$  per year, following a strong cooling phase in 1991–1993. The most intense warming event with  $\sim 2^{\circ}\text{C}$  increase in the SST took place during winters of the 1994–1996 period. It also coincides with  $4\text{ cm yr}^{-1}$  net sea level rise in the basin, and substantial change in the annual mean net fresh water flux from  $150\text{ km}^3\text{ yr}^{-1}$  in 1993 to  $420\text{ km}^3\text{ yr}^{-1}$  in 1997. The subsurface signature of warming is marked by a gradual depletion of the Cold Intermediate Layer (characterized by  $T < 8^{\circ}\text{C}$ ) throughout the basin during the same period. Winters of the warming phase are characterized by weaker vertical turbulent mixing and upwelling velocity, stronger stratification and, subsequently, reduced upward nutrient supply from the nutricline. From 1996 onward, the major late winter-early spring peak of the classical annual phytoplankton biomass structure observed prior to mid-90s was, therefore, either weakened or disappeared altogether depending on local meteorological and oceanographic conditions during each of these years. The effect of bottom-up limited unfavorable phytoplankton growth is reflected at higher trophic levels (e.g., mesozooplankton, gelatinous macrozooplankton, and pelagic fishes) in the form of their reduced stocks after 1995. *INDEX TERMS:* 1635

Global Change: Oceans (4203); 4215 Oceanography: General: Climate and interannual variability (3309); 4568 Oceanography: Physical: Turbulence, diffusion, and mixing processes; 4815 Oceanography:

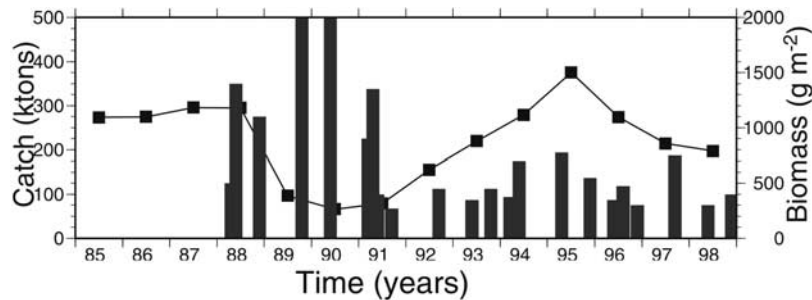
Biological and Chemical: Ecosystems, structure and dynamics; *KEYWORDS:* Black Sea, ecosystems, global warming, SST

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### 1. Introduction

[2] Over the last 3 decades, the Black Sea ecosystem was drawn into a highly unstable mode through a series of perturbations imposed by manipulation of river discharges, intense eutrophication introduced by exceptionally high amounts of nutrients and organic loads in river discharges, invasion by the gelatinous carnivore *Mnemiopsis leidyi*, and excessive fishing [Zaitsev and Mamaev, 1997; Oguz et al.,

2002a]. An order of magnitude increase in the primary productivity, dramatic reduction in the fodder mesozooplankton biomass, population outbursts of opportunistic species (e.g. *Noctiluca scintillans*), and gelatinous carnivore communities (e.g., jellyfish *Pleurobrachia* and *Aurelia aurita*, ctenophore *Mnemiopsis leidyi*) were the major modifications in the ecosystem structure during the period of maximum eutrophication from the mid-1970s up to the early 1990s [Shiganova, 1998; Vinogradov et al., 1999; Kideys et al., 2000; Oguz et al., 2001; Gucu, 2002; Kideys, 2002]. The period from 1988 to 1990, dominated by a massive outburst of the top predator *Mnemiopsis*



**Figure 1.** Annual Turkish anchovy catch (squares) in kilotons (the left axis), and total gelatinous macrozooplankton (the sum of *Mnemiopsis* and *Aurelia*) in g wet weight  $\text{m}^{-2}$  on the right axis (solid bars). The catch data are provided by the Turkish State Statistical Institute Reports. The gelatinous macrozooplankton data are reproduced from Figure 6 of *Kideys and Romanova* [2001].

population together with a collapse of fish stocks, constituted the most serious deterioration of the Black Sea ecosystem. Subsequently, as shown in Figure 1, while *Mnemiopsis* and *Aurelia* abundances reduced about three-fold in the early 1990s [Kideys and Romanova, 2001], observations suggested more diverse and abundant mesozooplankton community [Shiganova, 1998], and a gradual increase in small pelagic fish yield [Mutlu, 1999; Kideys et al., 2000].

[3] The recovery of the ecosystem observed during the first half of the 1990s was attributed to some protective measures imposed for controlling anthropogenic pollution, excessive nutrient loading and overfishing, as well as a weaker top-down grazing control of *Mnemiopsis* on mesozooplankton community. On the other hand, the Turkish anchovy catch data (Figure 1) as well as similar data for some planktivorous fish catches in Russia, Ukraine, and Georgia sectors of the Black Sea (see Figure 12 of Purcell et al. [2001]) show a systematic decreasing trend after 1995 even though the competition of gelatinous macrozooplankton (*Mnemiopsis* and *Aurelia*) community for food was no longer at a critical level for the pelagic fish groups (see Figure 1). The present study shows that similar adverse changes are in fact encountered at all trophic levels of the Black Sea food web from phytoplankton to fish during the second half of 1990s. These changes are shown to be triggered by the changes in the physical structure due to the climate-induced warming that took place in the Black Sea after 1993. The timing of this climatic signal coincided with abrupt increases in the sea level rise and the net annual mean fresh water flux in the Black Sea during mid-1990s as a consequence of the decadal scale climatic oscillations over the North Atlantic [Stanev and Peneva, 2002].

[4] The paper is organized as follows. Section 2 documents the decadal scale warming event as evident by more than  $2^{\circ}\text{C}$  rise in the annual mean sea surface temperature, and alterations in the physical properties of the upper layer water column. The subsequent modifications in different trophic levels of the food web are presented in section 3. A summary and conclusions are given in section 4. We note that our analysis is limited by the presentation of mainly basin-averaged data, since climatic warming imposes similar types of qualitative changes both in the physical and

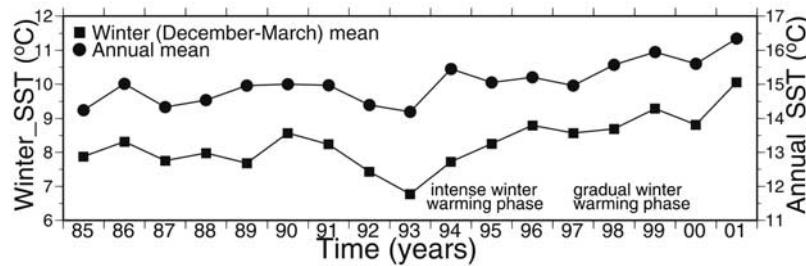
ecosystem structures of coastal and interior waters of the Black Sea

## 2. Physical Evidence for Decadal-Scale Warming

[5] The world ocean has been reported to warm by an average of  $0.31^{\circ}\text{C}$  between the surface and 300 m since the mid-1950s [Levitus et al., 2000]. Superimposed on the long-term trend, all major oceanic basins follow similar decadal scale variations introduced by changes in the structures of large-scale weather patterns. The sea surface temperature (SST) data, derived from a 9-km monthly, gridded NOAA/NASA AVHRR Oceans Pathfinder data set, provide a clear indication of warming of surface waters in the Black Sea during the 1990s. The warming is a robust signal of the system as shown by the basin-averaged, and monthly averaged data in section 2.1. It takes place over the entire basin with some spatial variability depending on the regional circulation and hydrological characteristics (section 2.2), and affects the upper layer temperature structure of the water column (section 2.3).

### 2.1. Temporal Variations of the Basin-Averaged SST

[6] The basin-averaged winter-mean (December–March) and annual-mean SST variations for the deep basin of the Black Sea excluding shelf areas shallower than 200 m during 1985–2001 period are shown in Figure 2. The winter-mean SST is  $8.1 \pm 0.3^{\circ}\text{C}$  from 1985 to 1991, and increases from its minimum value of  $6.8^{\circ}\text{C}$  in 1993 to a maximum of  $10.06^{\circ}\text{C}$  in 2001. The winter warming occurs in two phases; the 3-year period from 1994 to 1996 constitutes the “strong warming” phase characterized by  $\sim 2^{\circ}\text{C}$  rise in the basin-averaged winter SST. As will be shown in the next section, this major warming era promotes changes in the ecological regime of the Black Sea from 1996 onward. The period after 1996 characterizes the “gradual warming” phase in which the basin-averaged winter SSTs retained at least their 1997 level of warming. The warmer winter SSTs were correlated with milder winters characterized by weaker heat loss to the atmosphere and weaker wind stress forcing exerted on the sea surface [Nezlin, 2001]. Close correlation between the monthly variations of SSTs and air temperatures in the Black Sea during the 1990s have been shown by Krivosheya et al.



**Figure 2.** Basin-averaged, winter (December–March) mean (squares) and annual-mean (dots) AVHRR sea surface temperature distributions in the Black Sea from 1985 to 2001. They are obtained from 9-km monthly mean, gridded NOASS/NASA AVHRR Oceans Pathfinder data set. The averaging excludes the shelf areas shallower than 200 m. The temperature scale for the winter-mean data is given on the left, and for the annual-mean data on the right.

[2002]. The warming trend is also equally well-pronounced in the annual-mean data in the form of linear SST rise by about  $2^{\circ}\text{C}$  from 1993 to 2001 (Figure 2). The annual-mean SSTs were slightly less than  $15^{\circ}\text{C}$  before 1991, dropped to  $14.2^{\circ}\text{C}$  in 1993, and then tended to increase linearly afterwards at a rate of  $0.25^{\circ}\text{C}$  per year, with the highest annual mean value of  $\sim 16.4^{\circ}\text{C}$  measured during 2001.

[7] Further supporting evidence for changes in the upper layer physical structure comes from the sea level data. The basin-averaged sea level anomaly distribution from 1992 to 1997 (Figure 3) obtained from the Topex-Poseidon altimeter measurements [Stanev *et al.*, 2002] suggests well-pronounced changes in the seasonal signal. The winter sea level anomalies gradually vary from strongest negative values of 17 cm at the time of coldest winters, such as 1993, to warmer winters with weaker negative sea level anomalies such as 6 cm in 1996 and slightly positive values in 1997. The overall strong linear trend of approximately  $4\text{ cm yr}^{-1}$  from 1993 to 1997 agrees remarkably well with the intense winter warming signature detected by the basin averaged winter SST data. It is further correlated well with a similar trend of increase in the net annual mean fresh water flux (precipitation plus runoff minus evaporation) from  $150\text{ km}^3\text{ yr}^{-1}$  in 1993 to  $420\text{ km}^3\text{ yr}^{-1}$  in 1997 [see Stanev and Peneva, 2002, Figure 4]. These changes imply disintegration of the prevailing basin-wide cyclonic circulation cell [Stanev and Peneva, 2002; Korotaev *et al.*, 2003], and weakening of the associated upward motion within the interior part of the basin after 1995.

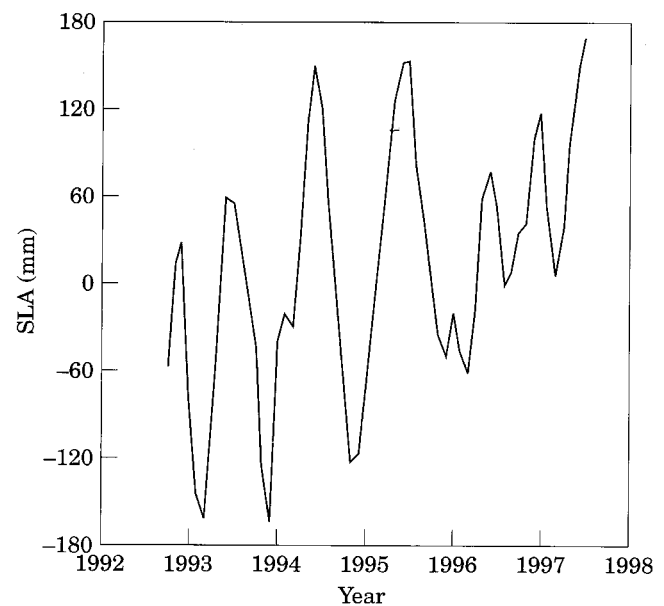
## 2.2. Regional Variability of Warming

[8] The essential features of the spatial variability of the winter warming is shown in Figure 4 using the winter-mean, December 2000 to March 2001 AVHRR SST distribution as an example. The temperature differences as much as  $3^{\circ}\text{C}$  are noted from the colder interior parts of the basin toward the warmer peripheral zone. In particular, regional meteorological conditions in the eastern part of the sea generally favor milder winters and warmer winter temperatures in the surface mixed layer. Consequently, the decadal warming signature is more evident in the eastern basin as well as around the periphery. The coldest part of the Black Sea lies within the narrow strip along the western coast receiving the highest freshwater discharge from Danube, Dniepr, and

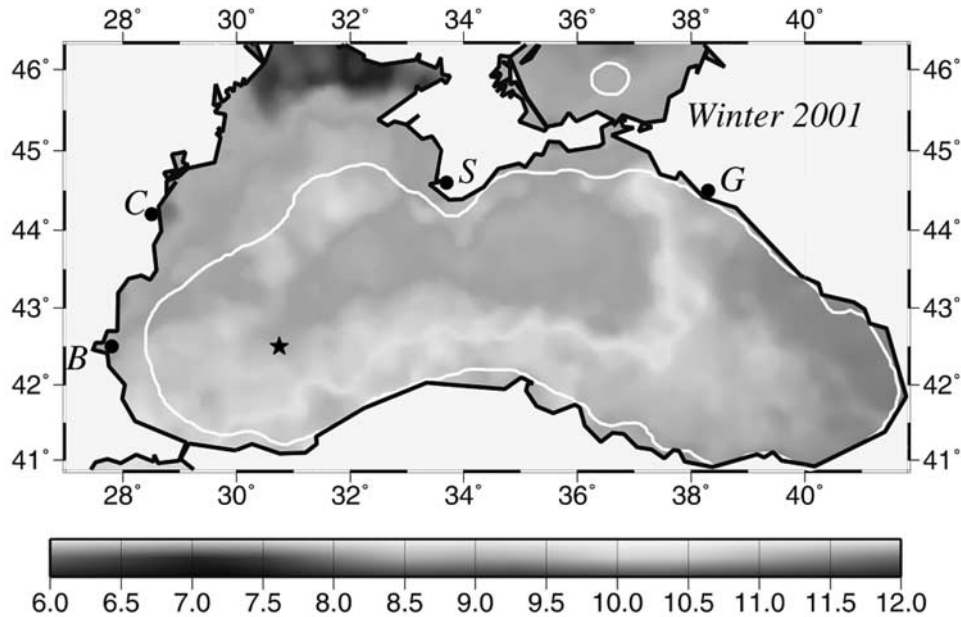
Dniestr Rivers. Even there, the mean winter SST of around  $3^{\circ}\text{C}$  during 1993 rose to  $\sim 7.5^{\circ}\text{C}$  in 2001. Such temperature differences between the eastern and western basins might constitute the main cause for the recent observed shift in the most abundant ichthyoplankton grounds to the southeastern Black Sea [Kideys *et al.*, 2000].

## 2.3. Subsurface Signature of Warming: Disappearance of the CIL

[9] The subsurface signature of the warming in the Black Sea can be traced from the structure of the Cold Intermediate Layer (CIL), characterized traditionally by temperatures colder than  $8^{\circ}\text{C}$ . This cold water mass, convectively generated every winter within the upper 50–75 m of the water column preserves its identity between the seasonal and permanent thermoclines during rest of the year. Vertical temperature profiles (Figure 5) taken at  $42^{\circ}30'\text{N}$ ,  $30^{\circ}45'\text{E}$



**Figure 3.** Seasonal distribution of basin-averaged sea level anomaly from 1992 to 1997 obtained from the Topex-Poseidon altimeter data [after Stanev *et al.*, 2002].



**Figure 4.** Black Sea winter-mean SST distribution for December 2000 to March 2001 period. The curve in white color around the periphery represents the 200-m topography contour separating the shelf and deep basin areas of the sea. Approximate locations of the Burgaz, Sevastopol Bays are indicated by the capital letters B and S, respectively. The locations of Constantza and Gelendzhik are shown by the letters C and G, respectively. The star symbol denotes the location of the vertical temperature profiles described in the text. See color version of this figure at back of this issue.

illustrate the temporal evolution of the CIL structure during the spring and summer months of 1993–2001. This site, located within the central part of the western basin cyclonic gyre shown by the star symbol in Figure 4, is typically characterized by coldest winter temperatures. The conditions shown there therefore reflect one of the most conservative subsurface warming signature with respect to coastal regions and the eastern basin.

[10] The April 1993 profile in Figure 5 shows a well-defined CIL with minimum temperatures of about 5.5°C extending to 50 m below the surface. The CIL is much thinner (around 20 m) and warmer with a core temperature of about 6.5°C during 1996. The CIL during 1999 is characterized by even warmer waters of about 7.5°C, and therefore maintains only a weak signature. During 2001, the CIL was not formed at all as a consequence of the minimum temperature in the water column staying above 8°C throughout the winter (see Figure 2). A narrow pocket of CIL with  $T \sim 7.8^\circ\text{C}$  at a depth of 50 m below the surface should be considered as a remnant of CIL from the previous years rather than a newly formed cold water mass. Further evidence on gradual depletion of the CIL during the second half of the 1990s in different parts of the sea is provided by *Staneva and Stanev* [2002] and *Krivosheya et al.* [2002].

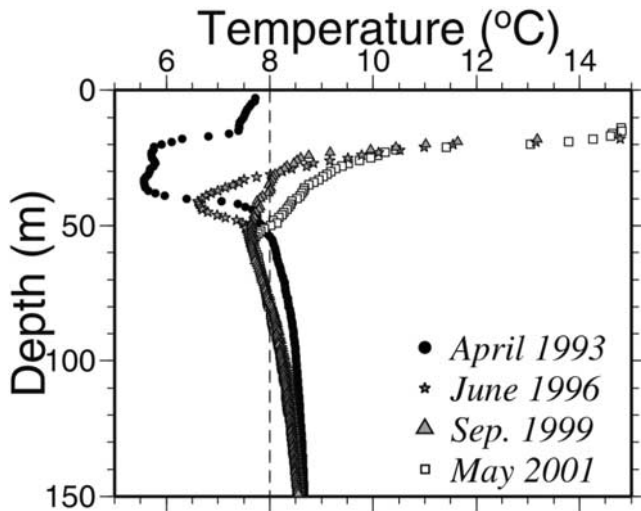
[11] An approximately 5–10 m rise of the anoxic interface during the second half of 1990s, reported by *Yakushev et al.* [2001], suggests a slight destabilization of the permanent pycnocline as a consequence of the warming of the surface waters. There is, however, no conclusive evidence for deeper penetration of warming to sub-pycnocline waters

below  $\sim 150$  m. The response of warming is mainly confined into the upper layer due the presence of strong density stratification, possibly except in regions of several deep and quasi-permanent anticyclonic eddies.

### 3. Biological Responses to Change in Marine Climate

#### 3.1. Phytoplankton Pigment Structure Prior to Mid-1990s

[12] The Black Sea ecosystem up to mid-1990s has been generally characterized by two distinct phytoplankton biomass peaks [*Vedernikov and Demidov*, 1997; *Yunev et al.*, 2002; *Oguz et al.*, 2001; *Sorokin*, 2002]. As inferred from the monthly mean surface chlorophyll concentrations (a proxy for phytoplankton biomass) in Figure 6, the major peak occurred either during February or March. It represents a typical early spring bloom event observed in temperate latitudes following nutrient accumulation in the euphotic zone under strong winter mixing conditions. The secondary peak took place in October or November, and coincided with the initiation of vertical mixing as well as autumn rebounds in gelatinous carnivore populations and subsequent decreases in herbivorous mesozooplankton stocks [*Oguz et al.*, 2000, 2001]. The surface mixed layer is characterized by much lower chlorophyll concentration during the warm part of the year, from May to September, due to limited nutrient supply into the mixed layer. An exceptional case of the surface intensified phytoplankton bloom event during June 1992 is marked in Figure 6 by



**Figure 5.** Vertical profiles of temperature within the uppermost 150-m water column measured at station 42°30'N and 30°47'E during April 1993 (shown by dots), June 1996 (stars), September 1999 (triangles), and May 2001(squares). The broken line shows 8°C temperature representing the upper and lower limits of the CIL.

chlorophyll concentration of  $\sim 2.0 \text{ mg m}^{-3}$  in the 1991–1995 data set. The lower part of the euphotic zone below the seasonal thermocline, on the other hand, can sustain some biological activity during this period [Oguz et al., 2001].

**3.2. Phytoplankton Pigment Structure After Mid-1990s**

**3.2.1. Observations From the Satellite Data**

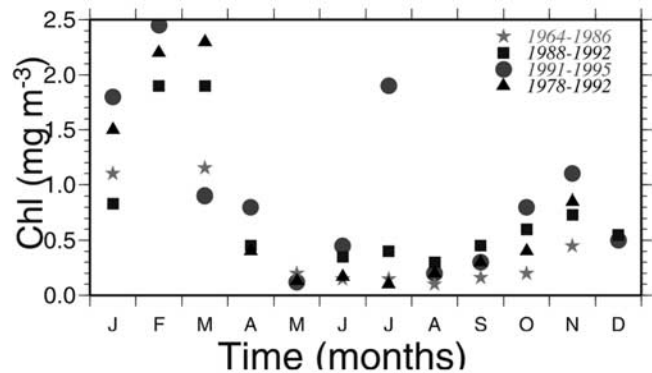
[13] Temporal variations of satellite-derived, basin-averaged monthly mean surface chlorophyll (Chl) concentration after November 1996 show a substantially different annual structure. This data set is formed by 9-km gridded SeaWiFS data for 1997–2002 (<http://seawifs.gsfc.nasa.gov/SEAWIFS.html>) and Ocean Color and Temperature Scanner (OCTS) data during November 1996 to June 1997 ([http://hdsn.eoc.nasda.go.jp/guide/satellite/sendata/octs\\_e.html](http://hdsn.eoc.nasda.go.jp/guide/satellite/sendata/octs_e.html)), and excludes the coastal regions shallower than 200 m. The SeaWiFS chlorophyll data are calibrated using the local algorithm [Suetin et al., 2001], which suggests an overestimation of the SeaWiFS chlorophyll concentrations by a factor of 2.

[14] The 1996–2002 surface chlorophyll time series data shown in Figure 7 reveal a linear seasonal trend of decreasing Chl concentrations from peak values in November to minima in July followed by a sharp increase from August to November every year. The phytoplankton production initiates every September, gradually intensifies and spreads over the basin in October, and finally attains its strongest phase in November [Oguz et al., 2002b]. The peak is particularly pronounced in 1996, 1998, 2000, and 2001 with values around  $0.9 \text{ mg m}^{-3}$ . The November peak was shifted to October in 1998, and was broader in autumn 1999. The autumn bloom episode generally terminates in January as indicated by gradual reduction

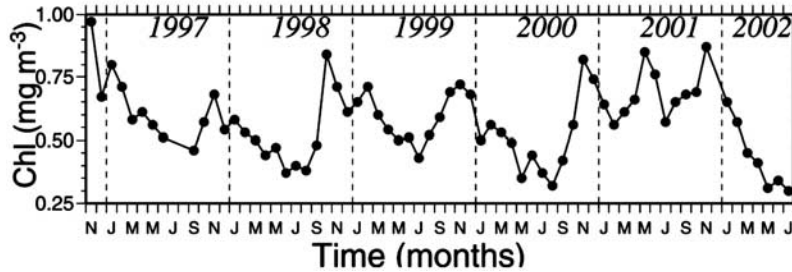
of chlorophyll concentrations to their minimum levels in winter months.

[15] The second major difference emerging from the new satellite chlorophyll data set is an almost continuous decreasing trend of concentrations during the winter and spring months. The strong chlorophyll signal, which was seen as the most robust feature of its annual structure every February or March in the former data set, now appears as only a slight increase in concentrations by about  $0.1\text{--}0.2 \text{ mg m}^{-3}$  either in January or February. Two exceptions occur in the last 2 years of the data set. In 2002, it does not exist at all. In 2001, the February concentration is comparable with those of the previous years, but it is followed by a gradual increase in the subsequent spring months. Therefore the February peak is seen as a minimum during 2001.

[16] Figure 7 also shows slightly elevated chlorophyll concentrations every May–June, the most pronounced of which occurs in 2001. Their magnitudes are found to be inversely correlated with the intensity of *Emiliania huxleyi* bloom events [Cokacar et al., 2001; T. Cokacar et al., Satellite-detected early summer coccolithophore blooms and their interannual variability in the Black Sea, submitted to *Deep Sea Research*, 2003) (hereinafter referred to as Cokacar et al., submitted manuscript, 2003)]. When the May–June chlorophyll concentrations are relatively low as observed during 1998, 2000, and 2002, *Emiliania* blooms identified by the SeaWiFS normalized water-leaving radiance data emerge as strong, season-long, well-defined basinwide events. In contrast, when *Emiliania* bloom activity had a spatially patchy character as in May–June 1999, the May–June chlorophyll concentrations attain somewhat higher value signifying a mixed species composition of the phytoplankton community, and contribution of other species groups to the bloom episode. In the absence of an *Emiliania* signature, as suggested by the analysis of normalized water leaving radiance data during May–June 2001, on the other hand, chlorophyll concentrations become



**Figure 6.** Monthly mean surface chlorophyll concentrations ( $\text{mg m}^{-3}$ ) composed from different measurements carried out in deep parts of the sea. The data shown by stars represents the period of 1964–1986, squares for 1988–1992, circles for 1991–1995, and triangles for 1978–1992. The latter data set is provided by Vedernikov and Demidov [1997], whereas the others are provided by Yunev et al. [2002].



**Figure 7.** Monthly mean, basin-average surface chlorophyll ( $\text{mg m}^{-3}$ ) distribution obtained by 9-km gridded SeaWiFS data for 1997–2002 and Ocean Color and Temperature Scanner (OCTS) data during November 1996 to June 1997. The basin-averaging excludes the coastal regions shallower than 200 m.

as high as their autumn values. This implies a major contribution of non-coccolithophorid species to this particular phytoplankton bloom. The conditions that control characteristics of the early summer phytoplankton bloom structure in the surface mixed layer remain to be determined by field measurements.

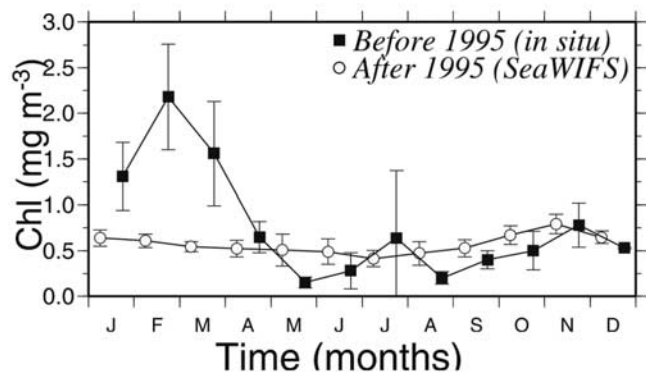
[17] The intimate relationship between climatic warming and the form of the annual chlorophyll structure is displayed in Figure 8 by expressing the data sets used earlier in Figures 6 and 7 in their multiyear average forms in the same plot. In the composite data set representing the monthly mean conditions for November 1996 to July 2002 (shown by dots in Figure 8), the small peaks given in the original data either each January or February are smeared out. In contrast to a well-pronounced peak of  $\sim 2.0 \text{ mg m}^{-3}$  in the data set prior to the mid-1990s, monthly chlorophyll concentrations remain steady at values of about  $0.5 \text{ mg m}^{-3}$  during the same period after the mid-1990s. The annual structure then acquires only a weak autumn peak of about  $0.75 \text{ mg m}^{-3}$  comparable to its counterpart in the former data set. It decreases gradually on its both side toward the summer and winter months.

### 3.2.2. Observations From Field Data

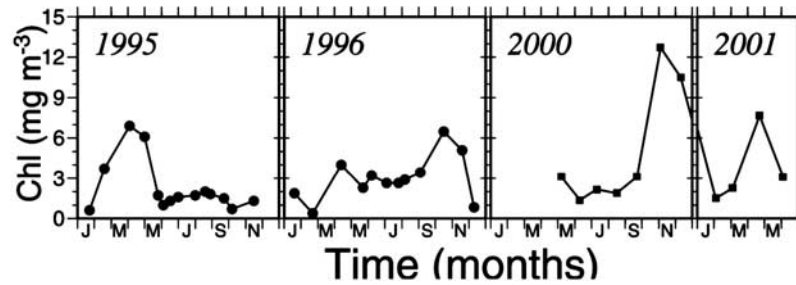
[18] In general, providing observational support for climate-induced changes in plankton populations is more difficult, and critically sensitive to sampling season and frequency, species community, location, as well as temporal and regional averaging of the data. Such problems often introduce major limitations for description and understanding of the ecosystem for deeper parts of the sea, where observations are more sparse and not particularly suitable for making reliable assessments on long-term variations. We therefore restrict our attention here on time series data collected systematically at several coastal locations. The in situ surface chlorophyll data from Burgaz Bay along the Bulgarian coast during 1995, 1996, 2000, and 2001 (Figure 9) provide an independent support for the change in the annual chlorophyll structure after 1995. The measurements from 1995 reveal an annual structure similar to that shown in Figure 6, with the strongest chlorophyll peak of about  $7.0 \text{ mg m}^{-3}$  during spring. The rest of the data set, on the other hand, suggests its new form as given by Figure 7 with the primary peaks of  $\sim 7.0 \text{ mg m}^{-3}$  and  $\sim 13.0 \text{ mg m}^{-3}$ , respectively, during late October of 1996 and early November of 2000. The relatively weaker spring peaks of

$\sim 4.4 \text{ mg m}^{-3}$  and  $\sim 7.5 \text{ mg m}^{-3}$ , respectively, occur during April 1996 and April 2001. We note that much higher chlorophyll concentrations in Burgaz Bay are typical for eutrophic coastal waters of the western Black Sea. The absence of the late winter-early spring phytoplankton bloom signature is also observed by weekly measurements in Romanian coastal waters near Constantza during 2000 and 2001 [Bodeanu *et al.*, 2002]. The phytoplankton abundance in February–March period remains under one million cells per liter representing the lower limit of a bloom population, except in a few instances in February 2000. Similar findings have also been supported by the measurements along the topographic slope zone (200–1500 m depth range) of the northeastern Black Sea. The June–September mean chlorophyll content of the surface mixed layer was found to decrease from its 1987–1992 averaged value of  $0.97 \text{ mg m}^{-3}$  to 1995–2000 averaged value of  $0.39 \text{ mg m}^{-3}$  with their respective standard deviations of 1.14 and  $0.10 \text{ mg m}^{-3}$  [Vedernikov and Demidov, 2002]. These field observations therefore suggests that the observed changes in the Black Sea annual chlorophyll structure, monitored by the monthly basin-averaged SeaWiFS data, is a very robust signature of the Black Sea ecosystem.

[19] Two different forms of the annual phytoplankton biomass structure depending on winter meteorological and oceanographic conditions have also been noted earlier by



**Figure 8.** Monthly mean surface chlorophyll ( $\text{mg m}^{-3}$ ) distributions obtained by averaging all the data shown in Figure 6 (squares) and in Figure 7 (dots).



**Figure 9.** Chlorophyll ( $\text{mg m}^{-3}$ ) distribution obtained by measurements at surface waters of Burgaz Bay along the Bulgarian coast Black Sea during 1995, 1996, and 2000–2001 periods.

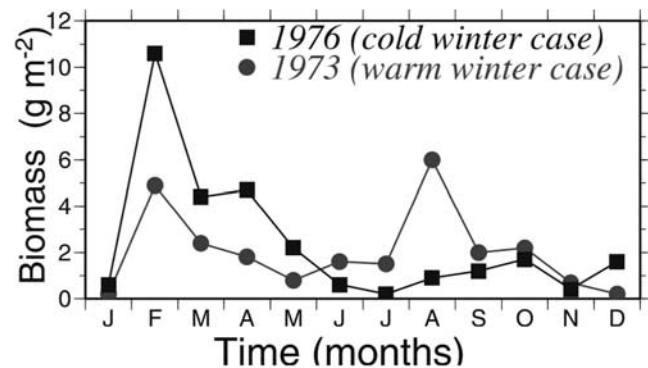
*Senicheva* [1980] on the basis of measurements from Sevastopol Bay in Crimea during two particular years in the 1970s (Figure 10). The year 1976 represents a standard cold winter case with the monthly mean winter SSTs of about  $4.0^{\circ}$  to  $6.8^{\circ}\text{C}$ , and with more intense vertical mixing and nutrient supply into the mixed layer during the winter season. The late winter-early spring phytoplankton bloom event was therefore the strongest of the year having the peak water column integrated biomass value of  $\sim 11 \text{ g m}^{-2}$ , followed by a secondary peak of about  $\sim 2 \text{ g m}^{-2}$  during the next autumn. This form resembles the structure shown earlier in Figure 6. In contrast, the winter of 1973 was considerably warmer with the monthly mean winter SSTs varying in the range from  $6.6^{\circ}$  to  $8.7^{\circ}\text{C}$ , and weaker mixing in the water column. The maximum phytoplankton biomass value of the late winter-early spring bloom event now constitutes only 40% of the former case whereas the autumn season reveals comparable biomass variations. A similar type correlation between intensity of winter meteorological and hydrodynamical conditions and the annual phytoplankton biomass distribution has also been pointed out by *Mikaelyan* [1997]. Apparently, the early 1970s was characterized by a strong warming trend [*Mikaelyan*, 1997; *Titov*, 2002] and sea level rise [*Stanev and Peneva*, 2002] as in the second half of the 1990s.

**3.3. Changes in Zooplankton Stocks After the Mid-1990s**

[20] Experience in different ecosystems suggested that abundance and distribution of certain key mesozooplankton species (e.g., the capepod *Calanus finmarchicus* in the sub-polar North Atlantic [*Heath et al.*, 2001]), are often correlated with concurrent changes in fish stocks, and therefore these mesozooplankton species establish the link between the amount of primary production by algae and the amount of fish available for harvesting. The measurements near Gelendzhik along the northeastern coast of the Black Sea (Figure 11) reveal two different forms of the annual mesozooplankton biomass distributions before and after 1995. Their distributions for the 1991–1993 period [*Khoroshilo and Lukasheva*, 1999] possess two distinct maxima with the primary one in March, and the secondary one in October–November. They, denoted in Figure 11a by P1 and P2, respectively, thus follow the spring and autumn phytoplankton blooms. These two peaks are connected with the relatively weaker summer mesozoo-

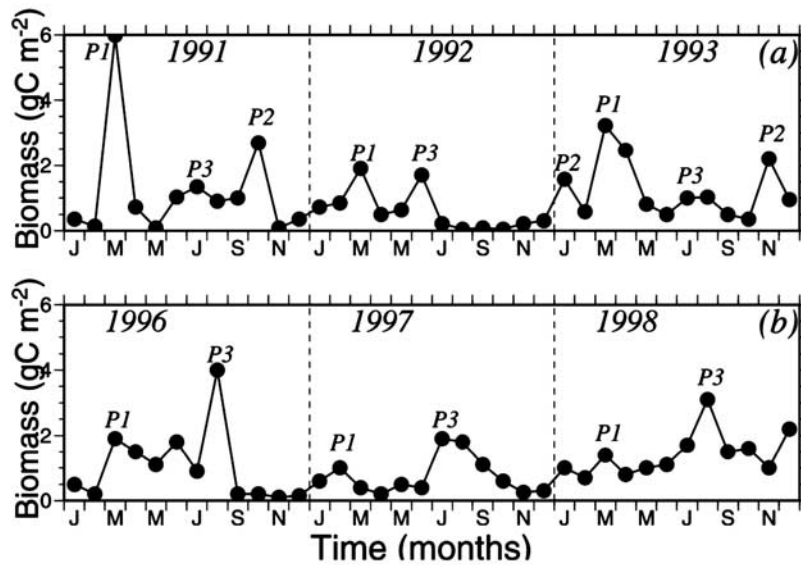
plankton activity (given by the peak P3). The intensity of the latter peak is controlled by two contrasting processes; the subsurface phytoplankton production, and the feeding pressure of anchovy larvae which imposes its strongest control during the summer. This structure slightly differs during 1992 by weaker spring peak P1, and shift of the autumn peak P2 toward the following early winter season. We recall that the 1991–1993 period corresponds to the coldest years of the last decade with minimal level of gelatinous carnivore stocks (see Figure 1) and therefore is characterized by their weakest predation pressure on mesozooplankton community.

[21] Similar measurements carried out at the same site during the 1996–1998 period [*Shushkina et al.*, 2000], i.e., in warming phase of the Black Sea, show a systematically different annual pattern. The early spring mesozooplankton bloom (i.e., the peak P1 in Figure 11b) is no longer a dominant feature of their annual structure due to the bottom-up resource limitations in the spring primary and secondary productions. The autumn mesozooplankton biomass distributions after the mid-1990s (i.e., the peaks P2 in Figure 11b) are also somewhat lower than those of the early 1990s displayed in Figure 11a. On the other hand, the summer peak P3 emerges as the most dominant feature of the annual mesozooplankton structure after the mid-1990s. It possibly



**Figure 10.** Monthly averaged water column-integrated phytoplankton biomass (in  $\text{g m}^{-2}$ ) distributions in Sevastopol Bay during 1976 (squares) and 1973 (solid circles) representing, respectively, the cold and warm winter cases. The original data are given by *Senicheva* [1980].





**Figure 11.** The distributions of mesozooplankton biomass ( $\text{gC m}^{-2}$ ) near Gelendzhik along the northeastern Black Sea. Measurements are carried out during the (a) 1991–1993 period and (b) 1996–1998 period. The original data are published by *Khoroshilov and Lukasheva* [1999] for Figure 11a, and by *Shuskina et al.* [2000] for Figure 11b.

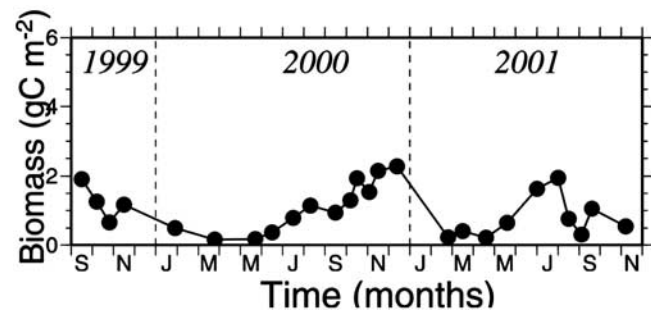
arises due to a relatively weaker top-down grazing pressure introduced by reduced anchovy population. As shown in Figure 12, somewhat similar annual structure is also observed within Sevastopol Bay from September 1999 to November 2001 [Finenko et al., 2003]. Once again, the mesozooplankton biomass is highest during summer months, even though the summer peak is shifted toward the autumn during the year 2000.

[22] The decreasing trends observed in both the gelatinous carnivore biomass and the anchovy catch data after the mid-1990s (Figure 1) imply that gelatinous carnivore and pelagic fish communities are in short supply of food by mesozooplankton prey. The gelatinous carnivore biomass after 1995 is almost comparable with their minimum level observed during the coldest years of the last decade (1991–1993), when they were unable to survive during exceptionally cold winter waters of about  $5^{\circ}\text{C}$  [Shiganova, 1998].

#### 4. Summary and Conclusions

[23] Changes in large scale atmospheric pressure and precipitation patterns are known to cause climatic variations both locally and in remote areas via teleconnection patterns. These interannual-to-decadal scale climate variations often appear in the form of the quasi-periodic sea level changes, warming and cooling cycles of the sea surface temperature, and impose significant impacts on ecology and ultimately on economical welfare of societies. All ecosystems respond rather gradually to external changes in climate, nutrient loading, biotic exploitation. However, under certain critical environmental conditions, gradual changes may be interrupted by more abrupt nonlinear-type changes in a stable state, or by a shift from one stable state to another [Scheffer et al., 2001]. Several examples are reported for the North Pacific by McGowan et al. [1998], Sugumoto and Tadokoro

[1998], Karl et al. [2001], Hunt and Stabeno [2002], for the North Atlantic by Reid et al. [1998, 2001], and Beaugrand et al. [2002]. Their causes are poorly understood, but it is believed that they are triggered mainly by changes in physical climate such as warming or cooling, and strong competition and predation among top grazers. For example, a major shift in north-northeast Atlantic marine ecosystems toward a warmer dynamical regime occurred after the mid-1980s. Warm water species in all copepod assemblages are extended northward more than  $10^{\circ}$  latitude at the expense of considerable reduction in cold water copepod species. These shifts have been linked primarily to the climatic warming of the Northern Hemisphere and to a lesser extent the winter North Atlantic Oscillation index (NAO). Similarly, apparent correlations between the recruitment of four main tropical tuna species and the interannual ENSO and decadal Pacific



**Figure 12.** Distribution of water column integrated mesozooplankton biomass ( $\text{gC m}^{-2}$ ) in Sevastopol Bay from September 1999 to November 2001. The original data are published by *Finenko et al.* [2003] and was provided to us by A. Kideys.

Decadal Oscillation signal have been noted in the tropical Pacific Ocean [Lehodey *et al.*, 1998]. Chavez *et al.* [2003] reported periodic shifts between anchovy and sardine dominated fishery regimes in the Pacific Ocean depending on its multidecadal warming-cooling cycles in response to large scale atmospheric and oceanic forcing.

[24] The results presented here also demonstrate a similar link between a decade-long climatic warming and changes in ecological characteristics of the Black Sea in 1990s. The warming period may well be teleconnected to changes in the NAO cycle, and the climatic warming trend of the Northern Hemisphere. Stanev and Peneva [2002] pointed out that the constant sea level rise of ~12 cm in the Black Sea from 1993 to 1996 is correlated with the increased net fresh water flux into the basin (i.e., the difference between precipitation plus river runoff and evaporation), which in turn is correlated with the dramatic decrease of the NAO index (from +2 to -2) during the same period. Analyzing 120 years of sea level data from a set of tide gauge stations around the Black Sea and the NAO index time series data for the same period, Stanev and Peneva [2002] noted a series of similar events, the most notable of which took place during 1939–1942 and 1970–1972. The loss of the spring 1973 phytoplankton peak in Figure 10 may possibly be connected to the changes in the physical climate occurred during the latter cycle.

[25] The positive impacts of climatic warming in the Black Sea are to (1) provide more suitable spawning and overwintering grounds for small pelagic fishes; the SST data suggest the eastern basin and the southern coast are the preferred regions, (2) contribute to successful invasion of the *Mnemiopsis* predator *Beroe* to the Black Sea in the late 1990s [Shushkina *et al.*, 2000], (3) promote more efficient growth of plankton communities by increasing their metabolic processes. However, the most notable effect of warming is to reduce nutrient supply from the nutricline due to weaker turbulent mixing and stronger stratification during mild winters of all these years [Nezlin, 2001]. Consequently, the late winter-early spring peak of the annual surface chlorophyll distribution is eroded by more than half after the mid-1990s. Such poor new production-based biological activity in February–March is followed by equally poor regenerated production during rest of the spring season. Mesozooplankton abundance also remains in low level during the same period. Moreover, because mild winters are characterized by relatively shallower mixed layer and weaker light limitation on primary production, the autumn blooms are often extended toward winter months. Mesozooplankton stocks are therefore not sufficiently strong to meet annual food demand by pelagic fish community at a steady level in the second half of the 1990s. As warming prevails longer, the continual loss of nutrients from the euphotic zone against their limited supply from subsurface levels exerts stronger bottom-up limitation. Consequently, both mesozooplankton and pelagic fish stocks gradually decline.

[26] Enhanced stratification and decreased inorganic nutrient availability are likely to alter phytoplankton community structure toward smaller cells with more enhanced nutrient uptake rates and light absorption characteristics as reported elsewhere [Karl *et al.*, 2001]. Increased coccolith-

ophorids contribution to annual phytoplankton community structure during the last decade (T. Cokacar *et al.*, submitted manuscript, 2003) might be a possible indication of such changes. The presence of more efficient calcium carbonate pump then releases of CO<sub>2</sub> into the water column, which in turn reduces biologically mediated CO<sub>2</sub> uptake demand from the atmosphere. Thus a closer look at plankton community structure might provide better understanding of current structure of the biogeochemical pump, and its future form when the present warming cycle is replaced by a cooling cycle. Unraveling impacts of climate-induced transformations in the Black Sea ecosystem and biogeochemistry, and predicting its future state from the fishery perspective require integrated observational and modeling efforts. Because it is a nearly enclosed basin and thus controlled environment, the Black Sea provides great opportunities for investigating many details of ecological regime change phenomena introduced by climatic and anthropogenic forcing mechanisms.

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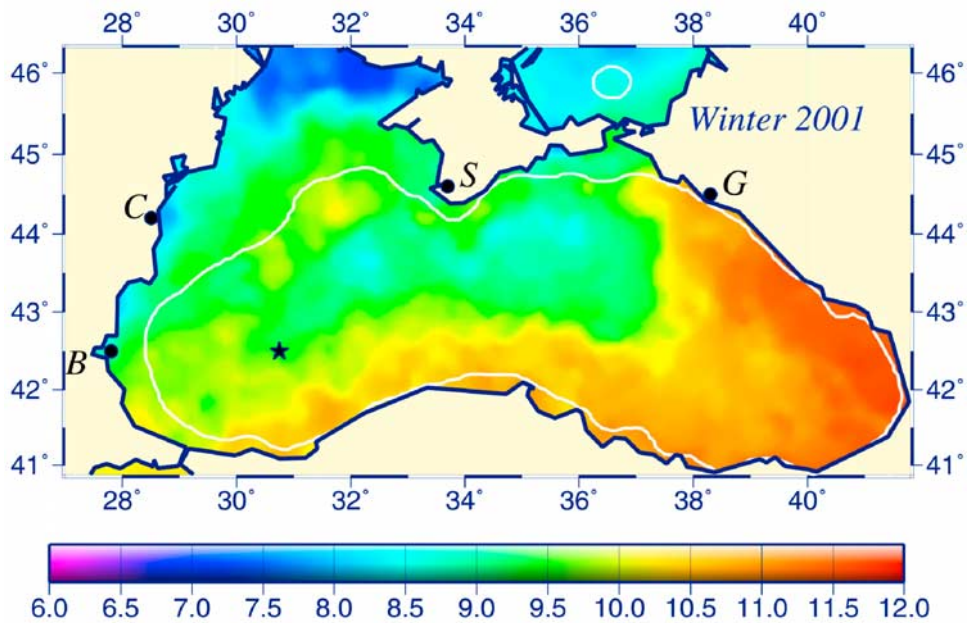
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**Figure 4.** Black Sea winter-mean SST distribution for December 2000 to March 2001 period. The curve in white color around the periphery represents the 200-m topography contour separating the shelf and deep basin areas of the sea. Approximate locations of the Burgaz, Sevastopol Bays are indicated by the capital letters B and S, respectively. The locations of Constantza and Gelendzhik are shown by the letters C and G, respectively. The star symbol denotes the location of the vertical temperature profiles described in the text.