Climatic Warming and the Decline of Zooplankton in the California Current

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Since 1951, the biomass of macrozooplankton in waters off southern California has decreased by 80 percent. During the same period, the surface layer warmed—by more than 1.5°C in some places—and the temperature difference across the thermocline increased. Increased stratification resulted in less lifting of the thermocline by wind-driven upwelling. A shallower source of upwelled waters provided less inorganic nutrient for new biological production and hence supported a smaller zooplankton population. Continued warming could lead to further decline of zooplankton.

Zooplankton play an important role in the biological cycling of carbon and other elements in the ocean. It is important to document trends in zooplankton biomass and also to understand how changes in climate may affect plankton. In this report, we describe 43 years of observations off the California coast and show that zooplankton have declined while the surface layers of the ocean have warmed. The decline is a major perturbation in the biota of the region because macrozooplankton form a significant part of the food web, may compete with larval fish for food, and are the main diet of some birds (*1*) and many schooling, commercially important fish species.

Our data come from the California Cooperative Fisheries Investigations (CalCOFI), a time series of physical, chemical, and biological measurements spanning an area of ocean greater than 130,000 km^2 (Fig. 1) since 1951. CalCOFI lines 90 and 80 are the most heavily sampled lines in the survey. Line 90 is in the Southern California Bight and line 80 is near Point Conception. Between 1951 and 1993 there were 222 cruises in which zooplankton were sampled by towing a 0.505-mm mesh net (*2*) at three or more stations (typically about 10) along line 90.

The zooplankton distribution is patchy in space and time. To produce a more Gaussian distribution for subsequent averaging, following the method described in (*3*), we computed the natural logarithm of the plankton volume, per 1000 m^3 of seawater strained, at each station. The transformed data were averaged over all stations from a given cruise along lines 90 and 80 to produce a time series for each line (Fig. 2, A and B). Zooplankton volumes decreased by about 80% from 1951 to 1993. In addition to this trend, there were low-frequency fluctuations with periods of years to decades. Because of the high interannual variability, it is uncertain whether the decline occurred gradually over the whole time series or more rapidly since the 1970s. The zooplankton volume at line 80 was consistently higher than at line 90, characteristic of the general northward increase in plankton volume, but the interannual variability and the downward trend were similar at the two lines.

We investigated the spatial dependence of the zooplankton decline by averaging over the initial and final 7-year periods of the survey at each station location (Fig. 3). These intervals were selected because they include a large number of cruises (71 and 27, respectively) but avoid effects of strong El Niño–Southern Oscillation episodes in California waters during 1958–59 and 1983–84. The decline from the initial to the final interval was about 80%. The difference appears to be uniform in space and is at least twice the standard deviation of the 7-year mean at each station. Spatial patterns were similar during both time intervals, with zooplankton increasing northward and shoreward.

For comparison with the zooplankton decrease, we analyzed temperature, salinity, and geostrophic transport in the upper 100 m at line 90. A warming trend previously identified (*4*) is evident (Fig. 2C). No trend was seen in upper-layer salinity or transport, although there were substantial decadal fluctuations in salinity, including a decrease of about 0.2 practical salinity unit (psu) from 1981 to the present.

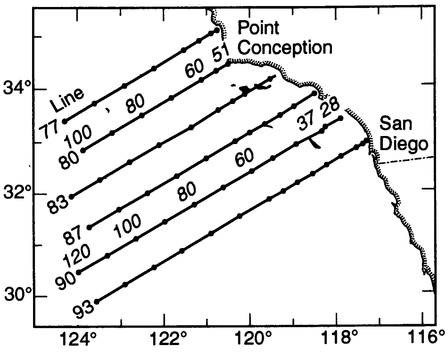

Fig. 1. CalCOFI survey plan, with station numbers (italics) indicated on lines 80 and 90.

We also averaged temperature, salinity, and geostrophic velocity over the same initial and final 7-year intervals as was done for the zooplankton data. Line 80 and line 90 surface temperatures (Fig. 4) warmed by an average of 1.2° and 1.6°C, respectively. Near Point Conception, the upwelling center for the region, surface temperature increased from 12.5° to 14.1°C between the two intervals. Temperature changes were reduced at depth, to 0.2°C at 180 m averaged along line 80 and at 270 m along line 90. As a result of the surface-intensified warming, the vertical stratification of the thermocline substantially increased and less cold waters came into contact with the atmosphere in the upwelling zones.

Salinity within the thermocline changed little between the initial and final periods. However, because of the warming, which in effect displaced isotherms downward, salinity on constant-temperature surfaces increased by up to 0.15 psu (at 10°C). Geostrophic velocity at line 90 showed an increase in the wind-driven recirculation—greater northward flow near shore and greater southward flow offshore—but no change in net transport.

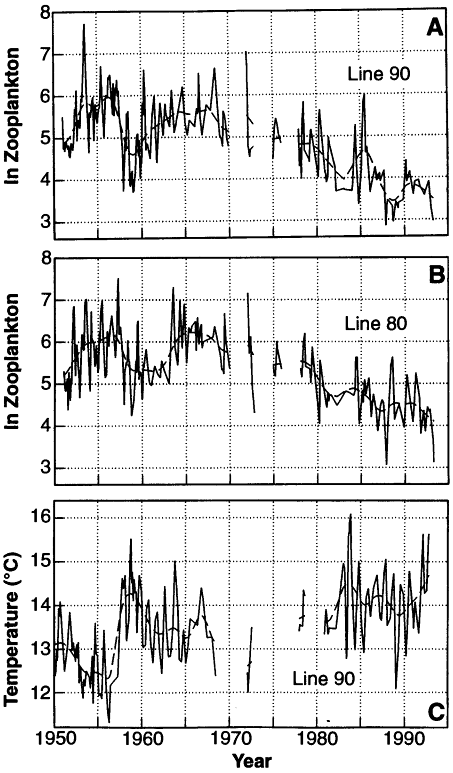

Fig. 2. Time series of log-transformed zooplankton volume (cubic centimeters of zooplankton volume per 1000 m^3 of seawater strained) for (**A**) line 90 and (**B**) line 80. On the logarithmic scale, a change of −1.6—which in (A) and (B) is the change from the mean of the 1950–1970 data to the minimum in the 1990s—is equivalent to an 80% decrease. (**C**) Time series of the upper 100 m average temperature for line 90.

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Bakun, analyzing a combination of directly estimated winds and onshore-offshore barometric pressure differences (geostrophic wind), has suggested that coastal winds favorable to upwelling off California increased substantially from 1946 to 1988 (5). However, there are possible biases in this climatology. The directly estimated winds reported by ships at sea tend to be irregularly distributed in space and time (5), and the geostrophic estimates may also be skewed because the upwelling zone is narrow compared to the spatially smoothed climatology. To illustrate, alongshore winds from CalCOFI stations during the period 1951 to 1993 (Fig. 5) show the narrow band of positive curl (upwelling) near shore at line 90 out to station 55, with negative curl (downwelling) farther seaward. While the reported increase in alongshore wind is consistent with the increase in oceanic recirculation noted above, its expected consequences are opposite the observed increase in temperature and decrease in zooplankton.

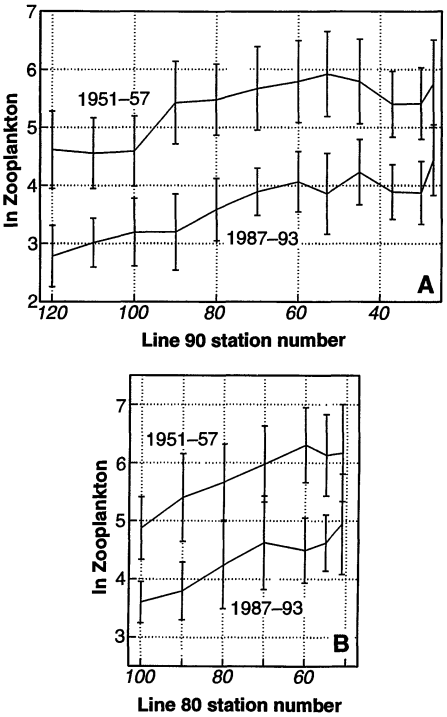


Fig. 3. Temporal average and standard deviation (vertical bars) of log-transformed zooplankton volume for all cruises during the periods 1951 to 1957 and 1987 to 1993 along (**A**) line 90 and (**B**) line 80. On this logarithmic scale, a change of −1.6 is equivalent to an 80% decrease. Station number units are equivalent to 4 nautical miles.

The observed changes in temperature and zooplankton over 43 years allow an assessment of thermal and ecological imbalances that are too small to measure instantaneously. A warming of the upper 100 m of the sea by 1°C requires a heat input of 4.3×10^8 J m^{-2}, equivalent to only 0.3 W m^{-2} for 43 years. Although such a small residual cannot be measured in instantaneous or annually averaged heat balances, the long-term warming is robust and has had substantial effects on coastal sea level rise (4).

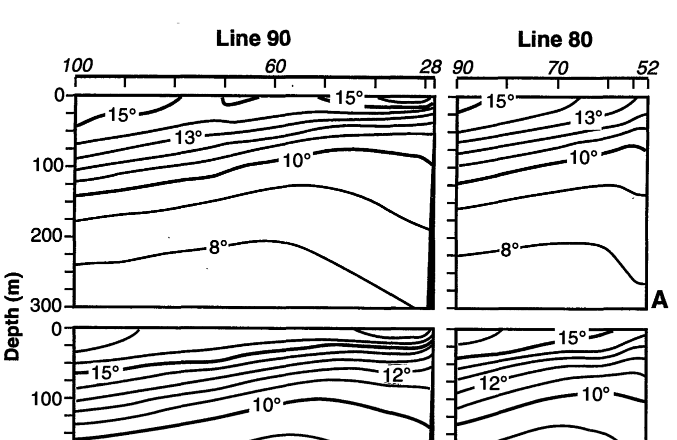
The observed rate of decrease in zooplankton volume is 6×10^{-10} s^{-1} for an 80% decline over 43 years. On long time scales, zooplankton biomass is controlled by the net effect of decreases due to excess mortality over reproduction versus increases through advection by ocean currents from the north (6). Advective input is estimated to be $\sim 5 \times 10^{-8}$ s^{-1}, far larger than the observed trend. The long-term trend thus is a small residual of much larger terms; it cannot be isolated by studies of advection, reproduction, or mortality. Nevertheless, it too is a dramatically large signal when accumulated over 43 years.

We suggest that the observed warming is linked to the zooplankton decline. As the sea surface is heated, the temperature difference across the thermocline increases, here by about 25% between the sea surface and a depth of 200 m. For a given alongshore wind stress, the (upwelling) displacement of the thermocline is inversely proportional to stratification (7). In other words, an increase in stratification results in reduced displacement of the thermocline. With less upward displacement, shallower layers bearing fewer nutrients are exposed to light, leading to less new production and ultimately to decreases in zooplankton. The mechanism is not a decrease in the volume of upwelled waters; it is a shoaling of the source of upwelled waters. Depending on relative magnitudes, this effect of heating could offset or even reverse the effect of an increase in wind stress.

The above mechanism relates a moderate surface warming to a major decline in the biota. In the waters along line 80, the nitrate concentrations decrease from about 30 μmol/liter at 8°C to 0 at about 14°C and warmer. Upwelled waters colder than 14°C contain nitrates but upwelled waters warmer than 14°C do not, having been stripped of nitrate at the sea surface in the basin-wide band of downwelling farther offshore. The effect of a sharper thermocline, with less vertical displacement due to wind stress, is to decrease the fraction of the year when wind stress is strong enough to lift nutrient-bearing waters to the sea surface near the coast. Thus, by insulating nutrient-bearing layers from the sea surface, a moderate degree of surface heating can greatly reduce the nutrient supply.

The observed trends in the California Current may be related to basin-scale changes in wind forcing. A strengthening of the North Pacific wintertime atmospheric circulation began in the late 1970s (8), near the time when the CalCOFI trends discussed here began to be clearly seen. A number of other effects have been observed in relation to this basin-scale change (9). It is possible that a shift in ocean circulation, such as at the bifurcation of the west wind drift reaching North America (10), might

Fig. 4. Average temperature (°C) for the periods 1951 to 1957 (**A**) and 1987 to 1993 (**B**) along line 90 (left) and line 80 (right).



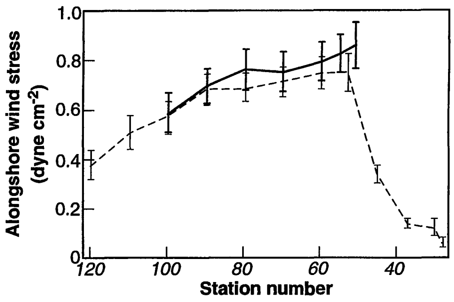


Fig. 5. Temporal mean and standard error (vertical bars) of alongshore equatorward wind stress (wind toward 150°) at CalCOFI stations along line 80 (solid) and line 90 (dashed) between 1951 and 1993, using stations with more than 80 observations.

import warmer water, decrease the supply of nutrients, or decrease the volume of zooplankton carried by the California Current.

The future consequences of the observed decline in zooplankton volume are closely tied to the question of causality. If the decline is part of a natural cycle that reverses in coming years, then any impact may be similarly transient. On the other hand, if the zooplankton decline is anthropogenic or is a natural trend of longer duration, then the large magnitude of the response is of great concern for the coastal ecosystem. The suppression of nutrient supply by enhanced stratification is not a mechanism confined to coastal oceans. If there is a global temperature rise of 1° to 2°C in the next 40 years and stratification increases globally, the biological impacts could be devastating. Our study also demonstrates that climate studies dominated by short-term process-oriented experiments cannot simply be extrapolated to decadal time scales, where the balance of terms is different from monthly or seasonal balances.

REFERENCES AND NOTES

1. A recent decline in a zooplankton-feeding seabird, the sooty shearwater (*Puffinus griseus*), in southern California waters was reported by R. Veit, J. McGowan, and P. Pyle, CalCOFI Annual Conference, 25 to 27 October 1994 (unpublished abstract, Marine Life Research Group, Scripps Institution of Oceanography, University of California, San Diego).
2. From 1949 until 1978, zooplankton were measured by oblique net tows with the use of a ring net (mouth diameter 1 m) and a towing bridle [P. Smith, *CalCOFI Atlas No. 20* (State of California, 1974)]. Target depths were 140 m from 1949 to 1969 and 210 m from 1969 to the present. Because zooplankton abundance decreases with depth, this change could have biased our post-1969 averages downward. The systematic effect of this change is not known, but it would be less than a 33% decrease in the post-1969 estimates even under the extreme assumption of no zooplankton below 140 m. After 1978, a "bongo" net was introduced that had been shown to collect more plankton per unit volume of water filtered, hence any systematic effect of this change would be an increase rather than the observed decrease in zooplankton abundance [J. A. McGowan and D. Brown, *Scripps Inst. Ref. 66-23* (1966); E. Brinton and A. W. Townsend, *CalCOFI Rep. vol. XXII* (1981)].
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11. We are grateful to the scientists and technicians responsible for the existence and high quality of the CalCOFI data. This study was supported by the Marine Life Research Group of Scripps Institution of Oceanography, University of California, San Diego, and by NSF grant OCE90-04230 (World Ocean Circulation Experiment).

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The Reaction of $ClONO_2$ with Submicrometer Sulfuric Acid Aerosol

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The measured reaction probability, γ, for the reaction of chlorine nitrate ($ClONO_2$) with 60 percent (by weight) sulfuric acid aerosol increases monotonically with particle size at 250 kelvin. The reacto-diffusive length (ℓ, the effective liquid depth over which reaction occurs) derived from these experiments is 0.037 ± 0.007 micrometer (95 percent confidence level for precision). The reaction probability for the reaction of $ClONO_2$ with 60 percent sulfuric acid aerosol doped with $\sim 7 \times 10^{-4}$ M hydrochloric acid at 250 kelvin is larger by about a factor of 4 than in the absence of hydrochloric acid and varies less with particle size (ℓ = 0.009 ± 0.005 micrometer). These results provide a test of the theory for gas-particle reactions and further insight into the reactivity of atmospheric aerosol.

Both $ClONO_2$ and HCl are relatively stable, gas-phase reservoirs of Cl in the stratosphere. Mechanisms that release Cl from these reservoirs enhance O_3 destruction by accelerating the Cl catalytic destruction cycles for O_3. The following heterogeneous reactions are important mechanisms for converting reservoir species into products that are readily photolyzed and that release Cl atoms:

$$ClONO_2 + H_2O \rightarrow HOCl + HNO_3 \quad (1)$$

$$ClONO_2 + HCl \rightarrow Cl_2 + HNO_3 \quad (2)$$

These reactions contribute to the winter-spring polar O_3 loss (*1*) and to the global destruction of O_3 during volcanic aerosol loadings (*2, 3*).

Although the uptake of $ClONO_2$ by bulk liquid H_2SO_4 solutions has been studied extensively (*4–6*), the reaction probability of $ClONO_2$ with submicrometer H_2SO_4 aerosol is unknown, and the parameters necessary to extrapolate the bulk results to the small aerosol that is characteristic of the stratosphere are uncertain. In this report, we describe measurements of the variation of the $ClONO_2$ reaction probability with particle size for submicrometer-sized H_2SO_4 aerosol, with and without HCl. These results provide an unprecedented test for the theory of the kinetics of gas-particle reactions and a basis for understanding the reactivity of atmospheric aerosol.

The rate of processing of a gas-phase reactant (x) by monodisperse aerosol is given by

$$\frac{d[x]}{dt} = -\gamma c \pi r^2 [P][x] \quad (3)$$

where γ is the reaction probability, c is the mean molecular speed of x in the gas phase, r is the particle radius, $[P]$ is the particle concentration, and $[x]$ is the concentration of x in the gas phase. For clarity, we do not discuss the small corrections for the diffusion of the gas-phase reactant to the particle (*7*). The reaction probability (γ) is given by

$$\frac{1}{\gamma} = \frac{1}{\alpha} + \frac{c}{4HRT\sqrt{k^I D_\ell}\left[\coth\left(\frac{r}{\ell}\right) - \frac{\ell}{r}\right]} \quad (4)$$

where α is the accommodation coefficient (the fraction of collisions that lead to accommodation by the liquid surface), H is the effective Henry's law coefficient (ratio of the liquid-phase concentration to the gas-phase concentration at equilibrium), R is Boltzmann's constant, T is the absolute temperature, k^I is the first-order rate coefficient for loss of x in the liquid phase, D_ℓ is the diffusion coefficient of species x in the liquid, and ℓ is the reacto-diffusive length defined as $\ell = \sqrt{D_\ell / k^I}$ (*3, 8, 9*). The reacto-diffusive length ℓ is the effective depth of liquid in which reaction occurs. Equation 4 demonstrates that the resistance to loss of the gas-phase species ($1/\gamma$) is the sum of the interfacial resistance ($1/\alpha$) and the reactive resistance (second term on the right side of Eq. 4). For small reaction probabilities ($\gamma \ll \alpha$), the reaction probability is given to a good approximation by

$$\gamma \approx \frac{4HRT\sqrt{k^I D_\ell}\left[\coth\left(\frac{r}{\ell}\right) - \frac{\ell}{r}\right]}{c} = \gamma_0\left[\coth\left(\frac{r}{\ell}\right) - \frac{\ell}{r}\right] \quad (5)$$

where γ_0 is the bulk reaction probability.

When ℓ is small relative to the size of the particle ($\ell \ll r$), the reactant is consumed

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