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Deep-water warming trend in Lake Malawi, East Africa

Abstract—We use historic water temperature measurements to define a deep-water warming trend in Lake Malawi, East Africa. Over the past six decades, the temperature of the deep water below 300 m has increased by ~0.7°C. The warming trend is due mainly to the reduction of cold-water deep convection over this period, which is associated with milder winters in the region. Despite deep-water warming, density stratification was maintained at depths below 100 m. The observed warming trend was interrupted at least twice by abyssal cooling events that were associated with the wettest years on record. We propose that rainfall and cool river inflow are critical factors that control thermal structure and the rate of deep-water recharge in this deep, tropical lake.

The thermal structure and mixing of the deep lakes in the East African Rift are driven by regional climate, which has influenced the physical and biogeochemical properties of these lakes over the distant and recent past (Johnson et al. 2002; O'Reilly et al. 2003; Verburg et al. 2003). A common feature of these deep, tropical lakes is their permanent, temperature-induced density stratification and their lack of complete convective mixing over at least the past century. Mechanisms that may lead to deep ventilation and cooling have been proposed, but direct observational evidence of these processes has remained limited to deep lakes in climatically temperate regions (Wüest et al. 2005). Under multidecadal steady-state temperature conditions, this permanent stratification can only be maintained if the deep water is cooled by localized cold-water convection (intrusions) followed by horizontal homogenization at depth, thereby compensating for the geothermal warming and for the downward heat input that results mainly from the wind-driven vertical turbulent mixing. Conceptually, the main mechanisms that can lead to a long-term, deep-water warming trend are enhanced vertical heat input through turbulent mixing and reduced convective cooling. These two processes operate on different timescales—while the former affects deep-water temperatures on timescales of decades due to the slow renewal of deep water, the latter occurs more rapidly and episodically. Our results suggest that reduced deep convection has resulted in warming of the lake's deep water.

Lake Malawi, also Lake Nyasa or Lake Niassa, the southernmost (9°S–15°S) of the large African Rift lakes, is ~700 m deep and is anoxic below a depth of ~220 m (Fig. 1). The main features of the catchment's complex climatology are a distinctly warm and wet austral summer (Dec–Apr) and a cool and dry winter (May–Aug), followed by ~3 warm and dry months, Sep–Nov (Torrance 1972). Most rainfall results from the seasonal southward excursion of the intertropical convergence zone and its associated regional air masses. In summer, Lake Malawi surface-water temperatures reach ~28°C, while in winter, they decrease to <24°C, when the surface mixed layer deepens to ~100 m as a result of the windier and cooler weather. Then the temperature difference between the surface and the lower part of the permanently stratified deep water decreases to <1°C (Wüest et al. 1996; Vollmer et al. 2002). Enhanced cooling has been observed occasionally at the southern end of the shallow (~100-m maximum depth), ~80-km-long southeastern arm of the lake (Eccles 1974) during multiple weeks in the cooler and windier season in connection with local upwelling and evaporative cooling by the strong and dry southerly wind, locally known as the Mwera.

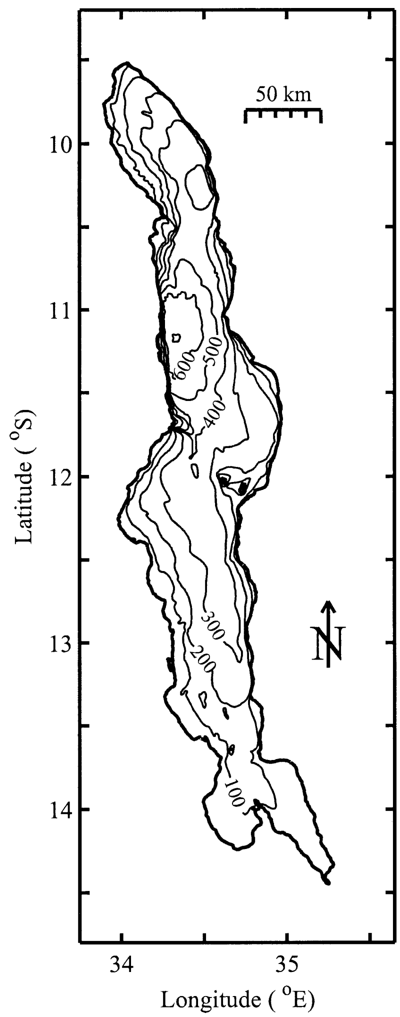

Fig. 1. Bathymetric map of Lake Malawi, East Africa, showing isobaths in 100-m intervals.

Methods—Deep-water temperature data (>300 m) from six expeditions previously unpublished in the open literature are based on measurements using reversing thermometers (1974, 1980, 1981, 1992) and on measurements using modern conductivity-temperature-depth (CTD) data (1995 and 1998/99; Vollmer 2002). An in situ thermometer comparison in 1974 by J.M.E. and D.H.E. revealed the necessity of correcting the 1964 (Eccles 1974) and 1976 (Gonfiantini et al.

1979) results by approximately −0.1°C (Vollmer 2002). All temperature profiles not reaching to maximum depth were extrapolated using vertical temperature gradients from a single profile collected in 1997 (Vollmer et al. 2002). Potential temperatures were calculated (Wüest et al. 1996) and, where possible, results were scaled to the International Temperature Scale ITS-90. The coarser resolution profiles were interpolated using spline-fitting techniques. The resulting profiles were then hypsographically averaged over the depth range of interest using bathymetric data provided by B. Halfman and T. Johnson (Large Lake Observatory, University of Duluth, Minnesota). For the shallow-water temperature trends (<300 m), the same procedures were applied. Temperature profiles not available in numerical form were reconstructed from their graphical representations. Large partially unpublished data sets for the 1990s were made available by G.P. and H.A.B. Overall uncertainties for the record are estimated at 0.02°C for the record before 1992 and about 0.01°C for the record thereafter.

We have used the University of East Anglia's Climatic Research Unit (CRU) 0.5°-resolution climate data set CRU TS 2.0 (New et al. 2000; Mitchell et al. pers. comm.) for the Lake Malawi catchment spanning 33°E–35°E and 9°S–15°S. The temperature data were normalized on a monthly basis to the mean temperatures of 1961–1990. In this data set, the mean number of stations that are in range of the grid boxes for the chosen region increased from 1940 (∼7) to 1960 (∼16) and decreased again from 1990 (∼17) to 1999 (∼8), with one abrupt change (∼4) in 1951. Based on this information and a comparison of the resulting temperatures with independent observations at Salima (34°24′E, 13°45′S) for 1980–1993 (Patterson and Kachinjika 1995), we conclude that the air temperature trends for 1940–1999 are not a computational artifact of the climate data set.

Results and discussion—Calculated volume-averaged temperatures for the deep waters of Lake Malawi over the past six decades are shown in Fig. 2a. Data from the literature (Bertram et al. 1942; Beauchamp 1953; Eccles 1962, 1964, 1974, 1988; Jackson et al. 1963; Gonfiantini et al. 1979; Halfman 1993; Vollmer et al. 2002) are supplemented by our previously unpublished data from 1974, 1980, 1992, 1995, and 1998/1999. Volume-averaged potential temperatures for the deep water (>300 m) increased from ∼22.0°C to ∼22.7°C between 1939 and 1999, at an overall rate of 0.10°C ± 0.008°C per decade. If we divide this temperature record into three sections, separated by the two cooling events and each reducing the temperature by ∼0.1°C, then the rate of increase is ∼0.18°C per decade for each section.

Lake-wide data sets were used to compile temperature records for shallower layers (Beauchamp 1953; Iles 1960; Eccles 1962, 1964; Jackson et al. 1963; Ferro 1977; Degnbol and Mapila 1982; Patterson and Kachinjika 1995; GP, HAB unpubl. data). The temperature at 100 m (typically the base of the mixed layer) as plotted in Fig. 2b, increased by ∼0.7°C (∼23.0°C to ∼23.7°C) at a rate of 0.06°C ± 0.02°C per decade, a change comparable with that of the deep water over the same six-decade interval. The data for the depth layer 10–50 m in the winter months of May–Aug are plotted in Fig. 2c and show a temperature increase of more than 1°C between 1940 and the late 1950s, followed by a smaller increase of 0.13°C ± 0.13°C per decade over the subsequent four decades. In contrast, the temperatures in the summer months of Dec–April plotted in Fig. 2d show little change over the entire record. These results indicate that the vertical density stratification over the full depth range has weakened during the summer months over the observed period. However, because of the comparable warming of the deep water and the water at the base of the mixed layer (∼100 m), the density stratification below 100 m has remained roughly constant over this period.

Historical wind-speed data for Lake Malawi are scarce. Measurements collected near the southeastern shore (Patterson and Kachinjika 1995) for 1980–1993 are suggestive of a slight decrease in wind speed over the later part of this record. However, National Center for Environmental Prediction reanalysis wind-speed data from the NOAA-CIRES Climate Diagnostics Center (http://www.cdc.noaa.gov) for 1948–2003 (925 mbar pressure level, 35°E, 10–15°S) show the absence of a trend in wind speed over the entire record. Patterns of wind direction have also remained unchanged except for 1958–1961. During this period, a major reversal occurred in the longitudinal direction from a predominantly southern component during the winters to a northern component during all seasons in these 3 yr.

If we assume no changes in mean wind stress over the period of the water-temperature observations, then the constancy of the deep-water temperature gradient suggests that the vertical turbulent heat transport has remained nearly constant over this period. Hence, the deep-water warming trend is most likely due to reduced cold-water convection. Assuming a vertical gradient of 1°C between 100 m and the deep water and a vertical turbulent diffusion coefficient of $3 \pm 2 \times 10^{-5}$ m^2 s^{-1} (Vollmer et al. 2000), we calculate a net yearly heat input to the water below 300 m of $1.5 \pm 1.0 \times 10^{17}$ J, which corresponds to a flux of $1.3 \pm 0.9 \times 10^{17}$ J m^{-2}. To this, we add a yearly geothermal heat input of 0.15×10^{17} J (Von Herzen and Vacquier 1967), resulting in a warming of the deep water of 0.25 ± 0.16°C per decade. This is in reasonable agreement with the observed warming rate of ∼0.18°C per decade (except for the intervals of cold convection) and implies that the heat removed by cold-water convection over this period was significantly less than these values.

Interannual regional climate variability is expected to result in fluctuations in the lake's thermal regime. Periods of milder winters with reduced convection, subsequent deep-water warming and reduced density stratification are likely to be interrupted by cooling events which, in the absence of long-term climate change, result in a re-setting of the lake's thermal structure. Such cooling events have occurred over the past six decades, but they were insufficient to reset the deep-water temperatures to the lower values found in the 1940s and are, therefore, suggestive of a trend in the atmospheric forcing over this period. Indeed, regional surface air temperatures (CRU TS 2.0 data set; New et al. 2000; Mitchell et al. pers. comm.) for the Lake Malawi catchment show an increase in the mean air temperature of the winter months Jun–Aug by ∼0.6°C over the last six decades (Fig. 2e), consistent with the deep-water warming trend being related to the generally milder winters in this region. The mean

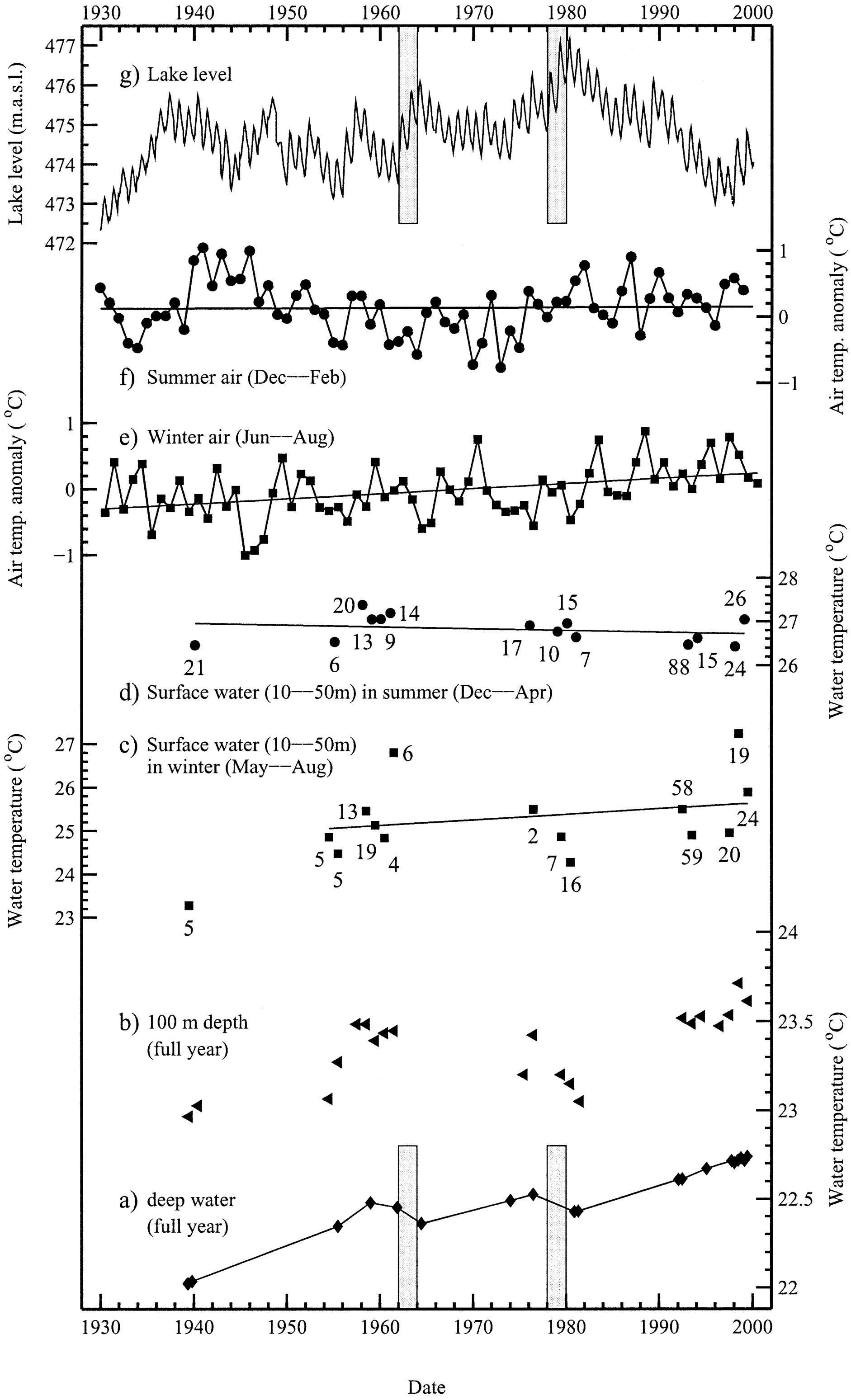

Fig. 2. Time series of Lake Malawi water temperatures, regional air temperatures, and lake levels for the period 1930–2000. Volumetrically averaged potential water temperatures are given in (a) for the deep water with depths greater than 300 m; in (b) for 100-m depth; in (c) for the depth range 10–50 m for the winter months, and in (d) for the summer months. The numbers adjacent to the symbols denote the number of temperature profiles that comprise each mean. Surface air temperature anomalies for the catchment of Lake Malawi as derived from the CRU TS 2.0 climate data set (New et al. 2000; Mitchell pers. comm.) are shown in (e) for the winter months, June–Aug, and in (f) for the summer months, Dec–Feb. Lake-level data are given in (g) and show seasonal variability and rise during the wet periods 1961–1963 and 1977–1979 (shaded areas).

surface air temperatures for the summer months Dec–Feb show some variability but are characterized by the absence of a trend over the past six decades (Fig. 2f), which is in agreement with the relatively stable summer water temperatures of the lake's surface waters over this period (Fig. 2d).

Prior to the observed deep-water temperature reductions (Fig. 2a), the catchment experienced two of the wettest climatic periods on record in 1961–1963 and 1977–1979. Heavy rains affected large parts of Africa and India in 1961 (Conway 2002), extending at least into 1962 in the Lake Malawi catchment, while the late 1970s heavy rains were spatially more limited to southeastern Africa. During both periods, the total mean river inflow to the lake was more than 25% above the long-term average (Kidd 1983) and the lake levels increased over the course of 2 consecutive years, reaching its maximum level in the century-long record in 1980 (Fig. 2g).

We speculate that enhanced cold river water inflow, heavy rainfall as a source of cooler water, and likely also reduced summer insolation because of cloudiness caused a reduced net heat input into the top ~100 m of the seasonally stratified water during the warm and wet part of the year, thereby allowing a faster erosion of the thermocline in the winter months followed by enhanced convective activity. For Lake Malawi, most of the inflowing river waters sink below the surface. This is mainly due to their lower temperatures (Jackson et al. 1963; Kingdon et al. 1999; McCullough 1999). Typically, river temperatures are lower by a few degrees Celsius in summer but some can be lower by up to 10°C in winter compared with the respective surface-water temperatures. The depth of the river injection depends also on the amount of suspended sediments and temperature, and has been identified at ~50 m during one summer (Halfman 1993; Halfman and Scholz 1993) for the largest river of the catchment. Also, recent observations on Lake Malawi show that river temperatures during high discharge after heavy rains may be lowered by ~2°C (McCullough pers. comm.). The relative amount of entrained warmer lake water is likely to be reduced during such events due to the local accumulation of river water near the inlet (Wüest pers. comm.). In addition, river flows persist into the cooling dry season during exceptionally rainy years (Kidd 1983), causing entrainment of relatively cool surface water. These conditions result in a greater intrusion depth and are assumed to have prevailed during the two rainy periods, 1961–1963 and 1977–1979.

The pre-1980 deep convection event resulted in a significant decrease in the deep-water nutrient concentrations observed in 1980/1981 compared with concentrations found in other years, as is shown for silica in Fig. 3. Under normal conditions, dissolved nutrient concentrations are very low in the surface water and increase strongly with depth as a result of the decomposition of sinking biogenic matter and the slow renewal of the deep water (Vollmer et al. 2002; Bootsma et al. 2003). In contrast, silica and phosphorus concentrations for 1980 and 1981 decreased with increasing depth below ~300 m. This indicates that the convectively sinking water masses carried the signature of the nutrient-depleted surface waters (Si ~ 15 μmol kg^{-1}, P ~ 0.2 μmol kg^{-1}; Vollmer 2002; Bootsma et al. 2003) rather than that of high-nutrient river discharge (typically Si ~ 280 μmol kg^{-1}, P ~ 1 μmol kg^{-1}; Hecky et al. 2003). With reference to the silica profile for 1974, extrapolation to maximum depth gives an estimated deficit of ~8×10^{10} mol Si for the deep water in 1981. A sinking of 440 km^3 of surface water, cooled to a mean temperature of 22.2°C, could account for both the deep-water silica deficit and for a temperature reduction of 0.1°C. In an alternative scenario, if we assume that, for the 2-yr period 1977–1979, the total winter (May–Oct) river inflow of 21 km^3 was dense enough to sink to below 300 m, it would have had to entrain 22 times its volume of surface water to account for the silica deficit. If we assume that the mean winter river temperature was 15°C, then the mean temperature of the entrained lake surface water had to be 22.4°C to account for a deep-water temperature reduction of 0.1°C. Both entrainment factor and total volume of sinking water in these scenarios appear high—they are mainly constrained by the silica budget.

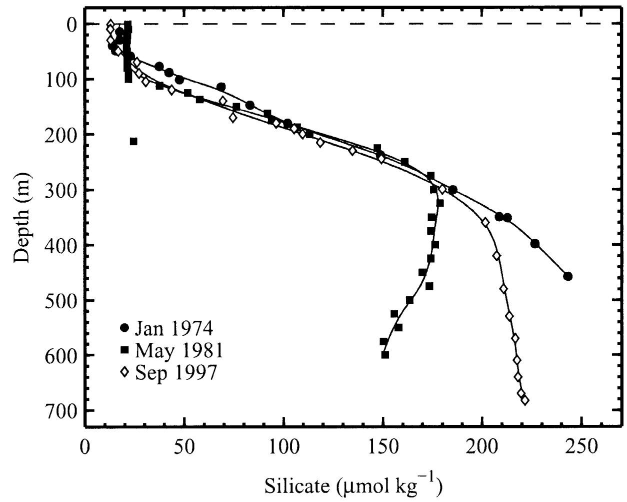

Fig. 3. Dissolved silica concentrations in Lake Malawi for 1974, 1981 (Vollmer 2002), and 1997 (Bootsma et al. 2003).

While deep-water warming of the neighboring Lake Tanganyika has been linked to reduced mixing caused by increased density stratification due to strong surface-water warming (O'Reilly et al. 2003; Verburg et al. 2003), the deep-water warming and potential reduction of the density stratification in Lake Malawi is suggested to be a result of reduced cold-water intrusions due to warmer winters. Sporadic cooling is caused by the sinking of cold water from the surface, possibly triggered by enhanced rainfall and river inflow. The more southerly Lake Malawi experiences greater seasonality in annual air temperatures and rainfall, and this greater seasonality may predispose Lake Malawi to greater convective cooling of deep waters compared with Lake Tanganyika. Also, while, for Lake Tanganyika, wind speed has significantly declined over the past decades (O'Reilly et al. 2003), there are indications that, for Lake Malawi, it has remained unchanged. Despite the similar morphometries of these two rift lakes, they may be sensitive to different modes of vertical mixing. These processes affect important biogeochemical properties of the lakes, including the concentration and vertical distribution of nutrients and dissolved oxygen. Future changes in the density stratification and the probability of an extreme overturn event may largely depend on the evolution of regional winter air temperatures and the reoccurrence of unusually wet periods.

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Diatom fatty acid biomarkers indicate recent growth rates in Antarctic krill

Abstract—We investigated the relationship between nutritional condition (levels of specific fatty acids) and growth increment (percentage growth per intermoult period, percentage IMP^{-1}) for Antarctic krill (*Euphausia superba*) collected from the vicinity of South Georgia in the austral summer 2002. There were correlations between percentage IMP^{-1} and the concentration (gram : gram dry weight) of the diatom biomarker fatty acids, 16 : 4(n-1) and 20 : 5(n-3) in tissues of individual krill, suggesting that the abundance of diatoms in the environment of the krill in the intermoult period prior to moulting was a key determinant of change in body length, a proxy for growth. This substantiates the view that diatoms are crucial for supporting high growth rates of krill, either as a direct food source or, indirectly, by enhancing production of microzooplankton and mesozooplankton based food webs.

Antarctic krill, *Euphausia superba,* are locally abundant in the Southern Ocean, where they are important for biogeochemical cycling and food web dynamics (Everson 2000; Atkinson et al. 2001). Understanding factors that control the growth and recruitment success of krill is an essential precursor to predictions of their distribution patterns and interannual variability. As in all crustaceans, growth rate in krill involves periodic moulting of the exoskeleton. Growth rate is thus a combination of moult frequency and the growth increment of moult. Quetin and Ross (1991) developed an instantaneous growth rate (IGR) technique to quantify krill growth rates that involves incubating animals individually for a few days after capture and, for those that moult, measuring the length increment at ecdysis. It is possible to calculate the growth increment, i.e., percentage growth per intermoult period (percentage IMP^{-1}), by measuring the difference in uropod length of both moult and animal. Application of the IGR technique during the Palmer long-term ecological research program (Ross et al. 2000; Quetin et al. 2003) indicated that highest growth rates of krill were associated with the latter stages of diatom blooms, whereas low growth was linked either to low phytoplankton biomass or blooms dominated by cryptophytes and prymnesiophytes.

Long-term, interannual, or high-resolution studies can be useful for characterizing the food environment of krill. However, such extensive coverage is rarely possible for open ocean biological oceanographers. Additionally, although it is relatively straightforward to characterize the potential food sources of krill, it is considerably more difficult to identify what krill are actually consuming. A marker within the krill that realistically reflects its food intake, both in terms of quantity and quality over at least an intermoult period (approximately 2–3 weeks in summer) and possibly longer is crucially required. Here we investigated the suitability of fatty acid biomarkers as a means of characterizing the nutritional condition of krill over intermediate timescales and how this influenced growth, as determined by the IGR technique. Fatty acids are increasingly used to identify specific