A 200,000-year, high-resolution record of diatom productivity and community makeup from Lake Baikal shows high correspondence to the marine oxygen-isotope record of climate change

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Abstract

   Siliceous microfossil succession was analyzed in a 200,000-yr sediment sequence recovered from the Buguldeika Saddle in Lake Baikal, Russia. Siliceous microfossil abundance varied among core depths from no preserved microfossils during inferred colder climate conditions to recent interglacial sediments containing over 300×10^6 microfossils per g dry sediment. Depth-age microfossil assemblage zones (CA-I to CA-IV) identified using correspondence analysis had high correspondence to stages in the marine $\delta^{18}O$ isotope record and could be partially aligned with Late Pleistocene glacial–interglacial cycling models from the mid-Siberian Highland. These observations suggest that Lake Baikal phytoplankton communities have responded to climatic changes driven by insolation parameters and global ice volumes on temporal scales similar to tropical and polar oceans. Microfossil zone CA-I (0–11.4 kyr B.P.) corresponded to the Holocene interglacial or $\delta^{18}O$ stage 1 (0–11.4 kyr B.P.), a period of higher production in Lake Baikal during a climatic optimum. Microfossil zone CA-II (12.3–18.7 kyr B.P.) corresponded to the Sartan glaciation and $\delta^{18}O$ stage 2 (12–24 kyr B.P.). Zone CA-III (21.3–73.2 kyr B.P.) comprised $\delta^{18}O$ stages 3 and 4. However, within zone CA-III, subzones CA-IIIa and IIIb (21.3–56.8 kyr B.P.) grouped were well aligned with $\delta^{18}O$ stage 3 and contained sediments deposited during the inferred Karginskiy interstade. Microfossil zone CA-IVa (77.4–129.2 kyr B.P.) and CA-IVb (130.3–172.5 kyr B.P.) included $\delta^{18}O$ stages 5 and 6, respectively, with sediments that were most likely deposited during the Taz glaciation and the Kazantsevo interstade. Climate-induced changes are reflected in production differences and in community composition specificity within microfossil zones or climate stages. This suggests that climate change drives major historical successional patterns in Lake Baikal's primary producer community; changes in primary producers must have further impacted the system's entire biota.

   Lake Baikal, the world's largest single body of fresh water by volume, is located in Central Asia (51°43′N, 103°44′E to 55°46′N, 109°37′E) along the Baikal Rift Zone between the Siberian Craton of the Eurasian continental plate and the Amur plate (Hutchinson et al. 1993; Moore et al. 1997). The lake is composed of three major basins separated by local highs in the lakebed. The Buguldeika Saddle-Selenga Delta separates the south from the central basin, and the Academician Ridge separates Baikal's deepest basin, the central, from the shallowest basin, the north (Fig. 1). Baikal is thought to have persisted as a lake for at least 25 million years, from the Miocene to present (Mats 1993). The large size, ancient age, and natural/economic resource potential of Lake Baikal have contributed to the uniqueness of its modern limnological conditions (Shimaraev et al. 1994), biota (Kozhov 1963), and environmental concerns (Galaziy 1980).

   The lake is situated in the mid-Siberian Highland, a wide geographical province within central Siberia (Arkhipov et al. 1986b). This region has been subjected to the impacts of glacial–interglacial cycling during the Pleistocene, as evidenced in pollen records and geomorphological features.

These impacts are known to have affected terrestrial communities surrounding Lake Baikal and should have affected biogeochemical processes within Lake Baikal. During the last 200 kyr, the mid-Siberian Highland has been influenced by three major glacial intervals: the Sartan glaciation (11–26 kyr B.P.), the Early Zyryanka (Murukta, 50–110 kyr B.P.), and the Taz glaciation from 130 to 180 kyr B.P. (Fig. 9; Arkhipov et al. 1986b; Bowen et al. 1986; Velitchko et al. 1989). Before and between these glacial stages were several interglacial or interstadial events. Modern Baikal is currently in the Holocene interglacial (0–11 kyr B.P.). The Sartan glaciation was separated from the Early Zyryanka by the Karginskiy Interstade (26–50 kyr B.P.), a period of alternating warm and cool periods. The Kazantsevo (Mikulino) interstade (110–130 kyr B.P.), a climatic optimum, separated the Taz glaciation from the Early Zyryanka. Preceding the Taz glaciation was the Shirta interstade (180–210 kyr B.P.). The Taz and Shirta were both within the more lengthy Bakhta glacial epoch (130–300 kyr B.P.).

   Because of its rift origin, latitude, and surrounding marginal mountain ranges, Baikal's lakebed is believed to have never been glaciated (Grosswald 1980; Arkhipov 1986a; Back et al. 1998, 1999). Glacial activity in the Baikal region has been limited to mountainous regions surrounding the lake (Back et al. 1998, 1999), and continental ice sheets have never scoured the region (Serebryanny 1984; Velitchko et al. 1989). Consequently, sediments have accumulated across large expanses of Lake Baikal's basins (Pampoura et al. 1993) since the Miocene to depths greater than 7.5 km (Hutchinson et al. 1993). These sediments preserve signals
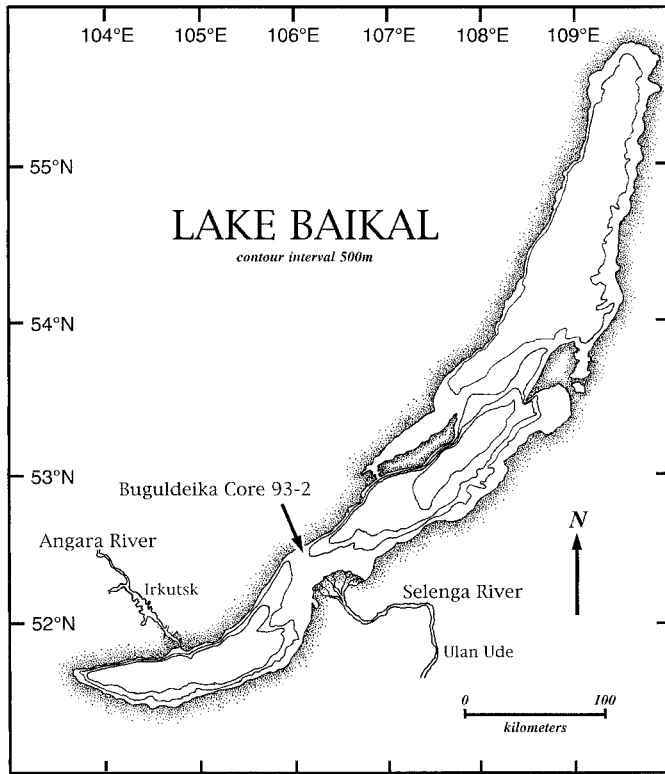

Fig. 1. Map of Baikal showing location of core site BDP 93-2 on the Buguldeika Saddle.

of changes that occurred both within and outside the lake basin. Furthermore, variations in sedimentation rates within and among basins (Edgington et al. 1991) provide researchers with opportunities to choose among scales of temporal resolution necessary for paleolimnologic and paleoclimatic reconstructions.

Several international scientific cooperatives have developed interest in sampling and interpreting portions of this lengthy sediment record. The Baikal Drilling Project (BDP), a cooperative of Russian, American, and Japanese scientists, was formed in 1989 with goals of reconstructing the geological evolution of the Baikal depression and reconstructing mid-Asian Quaternary climates (Kuzmin et al. 1993). During winter 1993, a barge was frozen in the ice over a local lakebed high, the Buguldieka Saddle (Fig. 1), a site identified from multichannel seismic readings (Hutchinson et al. 1993) as having a nonfaulted sedimentation history. Workers using a specialized drill rig modeled after the Ocean Drilling Project's (ODP) advanced hydraulic piston corer (APC) recovered two 100-m long cores (BDP-93 Baikal Drilling Project Members 1997). The first 1993 core, BDP 93-1, had only 72% recovery and was used for preliminary analyses by many research groups. The second core, BDP 93-2, had over 90% sediment recovery; the upper part of BDP 93-2 is treated in this paper.

Results from BDP 93-1 were synthesized in two preliminary reports (BDP-93 Working Group 1995; BDP-93 Baikal Drilling Project Members 1997) and indicated that numerous paleoclimatic proxies were identifiable in the Baikal sediment record. Notable among these proxies were the paleo-

magnetic measures, biogenic silica, pollen, clay minerals, and diatom stratigraphy. Core BDP 93-1 was dated using accelerated mass spectrometry (AMS) ^{14}C decay in total organic carbon and indicated a near-linear sedimentation rate during the last 25 kyr (Colman et al. 1996, 1999). Previous to this time period, paleomagnetic excursions in the core could be correlated to known excursions in the earth's magnetic field (e.g., the Biwa 1-Jamaica double excursion at 180 kyr B.P.) to create a date model for the lower core. With this date model, the core was believed to record approximately 500 kyr of Baikal history. A distinct change in sedimentation parameters was noted below 50 m by many analysts. Paleoclimatic signals were less resolvable, and it appeared that the lower section of the core was more influenced by sedimentation from the Buguldeika River system (BDP-93 Baikal Drilling Project Members 1997)

Paleomagnetic analyses suggested that there were higher rock magnetic concentrations during glacial times, which were inferred to result from a combination of less dilution by biogenic silica and depressed soil formation. These results were supported by changes in clay mineralogy inferred to represent differences in weathering patterns during warmer climates in both cores BDP 93-1 and BDP 93-2 (Yuretich et al. 1999). Paleomagnetic analysis was further used to determine the correlation between BDP 93-1 and BDP 93-2. Identical profiles were found, which suggests that direct date and depth correlations were appropriate (BDP-93 Baikal Drilling Project Members 1997; Colman et al. 1999).

Biogenic silica, composed mostly of diatom remains, has provided one of the most convincing proxies of past climate change in mid-Asia. Abundance of biogenic silica appears to reflect glacial–interglacial periodicity with greater abundance during interglacial times. Earlier work had suggested that this might be the case, as the Holocene climatic optimum (ca. 3–6 kyr B.P.) was preserved in shorter cores (<10 kyr B.P.) as a period of high diatom productivity and biogenic silica accumulation (Qiu et al. 1993; Carter and Colman 1994). The biogenic silica record in Baikal was also found to be strongly correlated to the marine δ^{18}O isotope record (Martinson et al. 1987) and, hence, correlated to the Milankovitch orbital parameters (Imbrie et al. 1984) by investigators working on various Baikal cores (Colman et al. 1995, 1999; Williams et al. 1997). Kashiwaya et al. (1997) generated similar Milankovitch periodicities using grain size and C:N ratios from BDP 93-1.

Biogenic silica peaks were well aligned with diatom abundance peaks in BDP 93-1 (BDP-93 Baikal Drilling Project Members 1997). Julius et al. (1997) analyzed 36 microfossil-bearing levels in core BDP 93-1 and identified 12 depths of microfossil abundance, noting peak abundances of diatoms during inferred interglacials. Each period of abundance had a characteristic diatom composition, with changes between zones inferred to be reflective of community and species response to climate forcing. Several of the peaks could be aligned with marine ^{18}O periods, and an initial record of past climates was proposed.

Following analysis of BDP 93-1, two conditions were noted that prioritized future analyses on BDP 93-2. First, change in sedimentation character below 50 m and apparent masking of paleoclimate signals before 250 kyr B.P. (BDP-

93 Baikal Drilling Project Members 1997) strongly encouraged finer scale analysis of the upper 50 m of BDP 93-2 (BDP-93 Baikal Drilling Project Members 1997; Yuretich et al. 1999). Second, the strong correlation between BDP cores 93-1 and 93-2, as evidenced by lithology and the paleomagnetic record, assured that preliminary results from BDP 93-1 could be applied to analyses on the more complete record in BDP 93-2. Toward these ends, sediments in the upper 30 m of core BDP 93-2 were sampled at a temporal resolution of ca. 1.0 kyr and analyzed for patterns in microfossil succession that reflect changes in the primary producer community of Lake Baikal and especially for changes that may be reflecting climatic impacts. A model consisting of AMS-radiocarbon dates and paleomagnetic alignment (Colman et al. 1999) dates this section of core from 0 to 200 kyr B.P.

Materials and methods

Core BDP 93-2 was recovered from Lake Baikal during March 1993 from a local lakebed elevation known as the Buguldeika Saddle (Fig. 1) in 354 m of water using a hydraulic piston coring system of aluminum drill pipe with 78-mm diameter plastic liners. Ninety percent recovery of the core's 92-m length was reported (BDP-93 Working Group 1995; BDP-93 Baikal Drilling Project Members 1997). Core sections, each approximately 1.5-m long, were sealed in their liners and transported to the U.S.A. for further subsampling.

For siliceous microfossil analysis, 1.0-cm subsamples were removed at approximately 16-cm intervals for the entire length of core BDP 93-2. Subsamples corresponding to these depths were also taken for biogenic silica analysis (Colman et al. 1999), whereas other researchers were provided samples for dating, paleomagnetism, lithological, and pollen analyses. Each subsample was freeze-dried to reduce microfossil breakage. Depending on earlier smear slide analysis of relative microfossil abundance, between 11 and 30 mg of freeze-dried sediment were prepared as in Stoermer et al. (1995) and Battarbee (1973). The four resultant coverslips with dried microfossils were mounted on microslides with Naphrax™.

Samples from 143 depths in BDP 93-2 between 7 and 2,870 cm were analyzed. These core depths cover from 0 to 200 kyr B.P. (Fig. 2a) as calculated from the dating model of Colman et al. (1999). The dating model recognizes two portions within the core. From 0 to 21 kyr B.P., 11 radiocarbon ages were calculated and indicated a near-linear sedimentation rate of 17.5 cm kyr^{-1} ($r^2 = 0.992$). Below this age (380 cm), dates were interpolated between five relative paleomagnetic intensity correlations at 566, 726, 910, 2,726, and 3,958 cm that corresponded to 39.6, 50.3, 65.1, 190, 264 kyr B.P., respectively (Colman et al. 1999). The use of interpolated near-linear sedimentation rates between control points is warranted because near-linear sedimentation had been shown in other Baikal cores. The phenomenon of linear sedimentation has occurred, even among glacial and interglacial periods, from a balance of in-lake biogenic and terrigenous sedimentation (Colman et al. 1995; BDP-93 Baikal Drilling Project Members 1997).

For siliceous microfossil analysis, a minimum of 500 microfossils was counted on a single transect on each of two replicate coverslips using brightfield oil immersion optics (N.A. > 1.32) capable of 1,200 × magnification. If 500 microfossils were not encountered in the first transect, an additional transect was counted on each of the two replicate coverslips. At some depths, counting 500 total microfossils was not possible due to extremely low abundance, and counting was terminated after two full transects. Diatoms were identified to the lowest taxonomic designation possible, and fragments of diatoms were mathematically reconstituted to whole microfossil (valve) units. Chrysophyte cyst morphotypes were enumerated using numerical identifiers that were consistent with previous studies (Edlund et al. 1995; Stoermer et al. 1995; Julius et al. 1997). Sponge spicules were uncommon in the Baikal material but were included in the counts. Raw counts were converted to absolute abundance and reported as millions of diatom valves or microfossils per g dry sediment. Total microfossil abundance, total planktonic diatoms, total benthic diatoms, and total chrysophyte cysts were also calculated and reported as absolute and relative abundance.

Eighteen diatom taxa and five chrysophyte cyst morphotypes were present at greater than 2% relative abundance in two or more core depths. Species-depth plots of the absolute abundance of each are presented and their absolute abundances formed the basis of ordination analyses. Matrices of core depths against species abundances were constructed and analyzed using correspondence analysis (CA) available in the ordination software CANOCO (Ter Braak 1988). Absolute abundances were log-transformed before analyses as outlined in Ter Braak (1988). Core depths that did not contain any microfossils were removed from ordination analysis.

Results

Over the last 200 kyr, total microfossil abundance in BDP 93-2 varied (Fig. 2b); some sample depths did not contain any microfossils (1,009 cm, 2,054–2,070 cm, 2,179–2,275 cm, 2,510–2,870 cm), whereas near-surface samples had values greater than 300 × 10^6 microfossils per g dry sediment. Ten peaks in abundance were distinguishable in total microfossil abundance (Fig. 2b), and these peaks corresponded to similar periods of biogenic silica accumulation (Fig. 2c). Weight percent biogenic silica values (clay-corrected, *see* Carter and Colman 1994; Colman et al. 1999) ranged from near zero to over 40% (Fig. 2b). The biogenic silica profile was distinguished by three major periods of accumulation (Fig. 2c; 0–250 cm, 580–900 cm, and 1,086–1,650 cm).

Three types of siliceous microfossils were encountered. Sponge spicules were found at most depths in very low abundance (<100,000 per g dry sediment). Diatoms were the predominant microfossil in the assemblages and contributed >80% relative abundance at most sample depths. The only levels that total diatoms contributed <50% to the assemblage were sampling depths with few or no microfossils. Diatom remains in Baikal can be separated into two habitat groups: planktonic and benthic. Planktonic taxa dominated the diatom assemblage at most depths (Fig. 3a,b). Benthic taxa were present in low relative abundance (<10%)
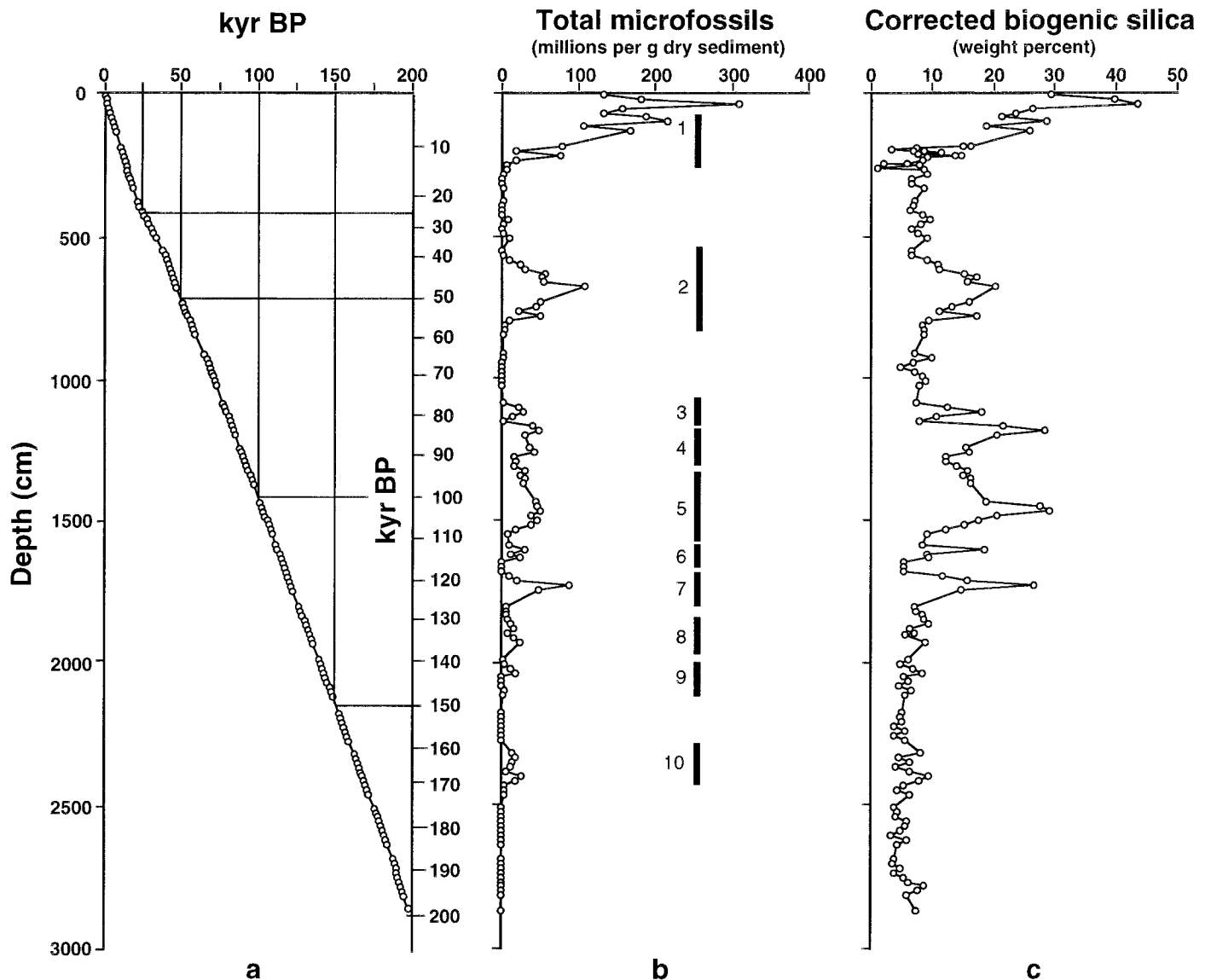
Fig. 2. Core BDP 93-2 (a) date model from Colman et al. (1999). (b) Absolute abundance (millions of microfossils per g dry sediment) of microfossils; bars represent numbered peaks of microfossil abundance. (c) Clay-corrected weight percent biogenic silica (percent weight of dry sediment; data from Colman et al. 1999).

throughout much of the core (Fig. 3c,d). Benthic diatoms were >20% of the total assemblage at only four depths (300, 474, 1,150, and 1,682 cm). These depths corresponded to periods of low microfossil and biogenic silica accumulation (Fig. 2c). Chrysophyte cysts are the final component of siliceous microfossil assemblages. Over 140 cyst morphotypes have been identified from Lake Baikal in our analyses. Two major peaks in cyst absolute abundance were apparent (Fig. 3e,f; near 85 and 1,746 cm). Cysts contributed >20% relative abundance at only five depths (474, 567, 1,025, 1,682, and 1,746 cm) that generally corresponded to periods of low microfossil accumulation.

Correspondence analysis of absolute abundances of 23 siliceous microfossil taxa across sampling depths proved explanatory for deciphering patterns of microfossil abundance and community makeup; eigenvalues and cumulative vari-ance of the species data on the eigenvectors are given in Fig. 4a,b. Correspondence analysis is an exploratory technique that assists in generating hypotheses to explain species abundance across sampling sites or core depths. Correspondence analysis (CA) uses iterative reciprocal averaging between sample and species scores to produce eigenvectors (axes) with species ordination scores as averages of sample ordinations and vice versa (Randerson 1993). When species and sample scores are displayed along the new ordination axes, their distribution may be used to explain patterns among the data.

Species and depth eigenvalues are plotted against CA axes 1 and 2 in Figs. 4a and 4b, respectively. Sampling depths and species scores are distributed along CA axis 1 according to core depth, and secondarily by species stratigraphy. Species to the left of the origin are exclusively found in the
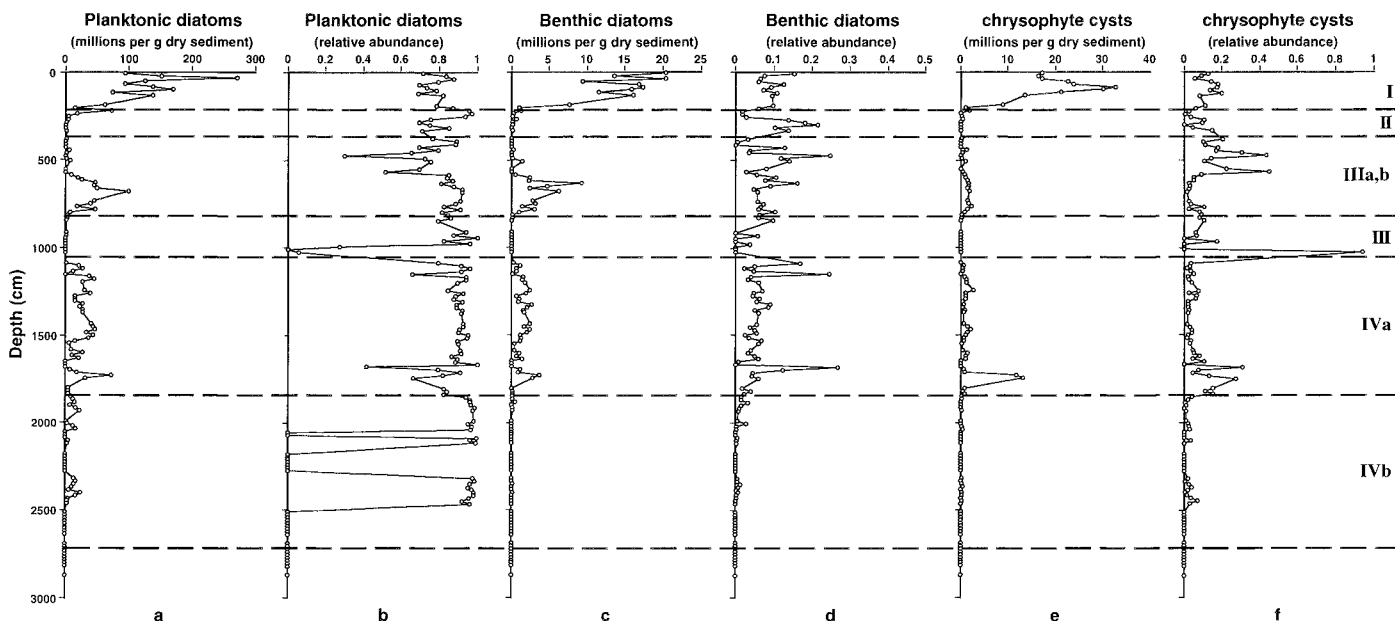
Fig. 3.   Absolute (millions of microfossils per g dry sediment) and relative abundance (relative to total microfossil assemblage) of siliceous microfossils, respectively, of (a, b) planktonic diatoms, (c, d) benthic diatoms, (e, f) total chrysophyte cysts.

shallower depths of the core, whereas species to the right of the origin are distributed farther downcore. Species grouped near the origin were distributed throughout the core (Fig. 4a). Sampling depth scores correspond to this pattern also, with deeper samples grouped right of origin and shallower samples grouped left of origin. The ordination of species and depth scores on the second CA axis is somewhat less clear. Samples appear ordinated by both abundance and species composition along axis 2, with samples of greater abundance placed below the origin and samples with lesser abundance above the origin. Samples are secondarily grouped with taxa that contribute most strongly to their absolute abundance. The ordination grouped samples into four major microfossil age-depth zones that are indicated in the ordination diagram (Fig. 4b). CA-I contains samples from 0 to 204 cm and CA-II has samples from 220 to 332 cm. CA-III groups samples 378–1,025 cm and contains two subgroups, CA-IIIa and CA-IIIb, grouping depths 378–679 cm and 732–812 cm, respectively. CA-IV contains samples from 1,086 to 2,464 cm and can be further subdivided into two subgroups with more recent samples located left of center in this group (CA-IVa) and deeper samples located right of this groups center (CA-IVb). CA-IVa contains samples from 1,086 to 1,836 cm (except 1,150 cm) and CA-IVb contains samples from 1,852 to 2,464 cm (except 2,102, 2,432, 2,448 cm). Two sample depths, 913 and 929 cm, were grouped as outliers within CA-IV. These two depths had low microfossil abundances and represented the uppermost distribution of *Stephanodiscus grandis,* a taxon more characteristic of CA-IV. Using the age-depth model (Colman et al. 1999), the following age limits on the microfossil zones are proposed: CA-I, 0–11.4 kyr B.P.; CA-II, 12.3–18.7 kyr B.P.; CA-III, 21.3–73.2 kyr B.P.; CA-IIIa, 21.3–47.2 kyr B.P.; CA-IIIb, 50.5–56.8 kyr B.P.; CA-IV, 77.4–172.5 kyr B.P.; CA-IVa, 77.4–129.2 kyr B.P.; CA-IVb, 130.3–172.5 kyr B.P. (Fig. 9).

The distribution and ecology of specific taxa in BDP 93-2 will be presented, especially as the distributions are related to the CA groups. Some taxa are distributed within specific CA zones, whereas others have been present in Lake Baikal during the entire time span of the core.

Aulacoseira baicalensis (K.I.Mey.) Simonsen is a Baikal endemic that is a major contributor to modern phytoplankton biomass during late winter to early summer. Strong development of *A. baicalensis* produces what are known as "*Melosira*-years" (*Melosira* is a former generic assignment for this taxon) in modern Lake Baikal (Kozhov 1963). *A. baicalensis* often blooms below ice cover in late winter (Skabitschewsky 1929), maintained in the water column by thermal mixing (Matthews and Heaney 1987; Bondarenko et al. 1993). In core BDP 93-2 it is distributed in the upper 17 m with five absolute abundance peaks (Fig. 5a) and was the dominant taxon in many levels of the core.

Cyclotella minuta (Skvortsov) Antipova, *Cyclotella ornata* (Skvortsov) Flower, and *Cyclotella baicalensis* Skvortsov and K.I.Mey. are three closely related taxa and may represent a historically related planktonic species flock that has evolved in Lake Baikal (Julius et al. 1997; Edlund 1998). These three species are generally considered summer and late-fall taxa (Kozhov 1963; Bradbury et al. 1994); however, recent research suggests that *C. ornata* and *C. baicalensis* are also common in pelagic waters during spring in "non-*Melosira*-years" (Likhoshway et al. 1996). *C. minuta* is distributed throughout the microfossil-bearing sediment layers in BDP 93-2 with 10 major absolute abundance peaks, notably in zones CA-I, IIIa, IIIb, and IV (Fig. 5b). *C. minuta* was found in high relative abundance throughout most sampled depths. *C. ornata* had an identical downcore pattern of distribution as *C. minuta* but was typically found in lower abundance (Fig. 5c). *C. baicalensis* was distributed only in the upper 16 m of BDP 93-2 with three absolute abundance
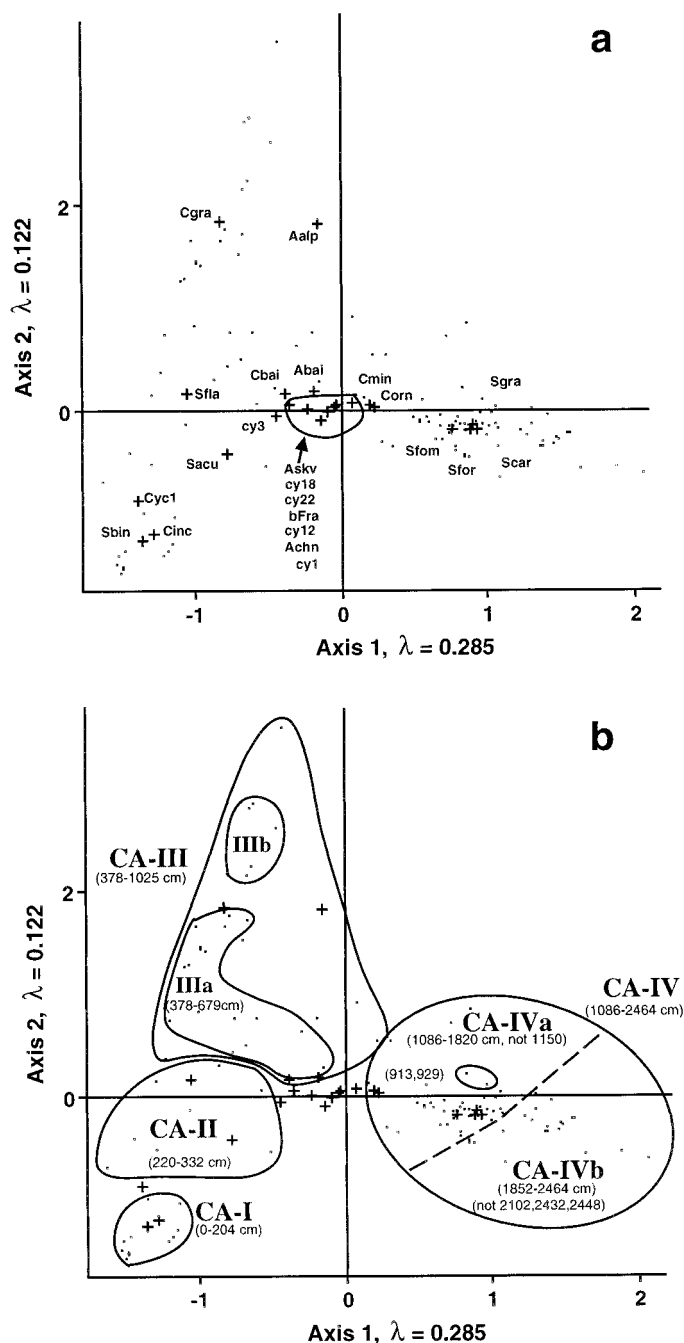
Fig. 4. Correspondence analysis ordinations of depth and species eigenvalues on axes 1 and 2 for absolute abundances of 23 predominant taxa in core BDP 93-2. (a) Ordination location of species identified by crosses and taxa codes, (b) microfossil-depth zones and subzones and their corresponding depths are indicated (CA-I to CA-IV).

maxima (Fig. 5d). Of the three *Cyclotella* species in this group, *C. baicalensis* is the largest in size but is also least abundant in the core.

Aulacoseira skvortzowii Edlund, Stoermer, and Taylor has been formally recognized only since 1996, when it was described by Edlund et al. (1996). This taxon was earlier reported from Baikal plankton and sediment collections as *Me-*

losira islandica ssp. *helvetica* O.Müll. (Kozhov 1963; Kozhova et al. 1982), *Aulacoseira islandica* (O.Müll.) Simonsen (Genkal and Popovskaya 1990*b*, 1991; Bradbury et al. 1994; Flower et al. 1995; Babanazarova et al. 1996), *A. islandica* ssp. *helvetica* (O.Müll.) Simonsen (Bondarenko et al. 1993), *A. islandica* var. *helvetica* nom. nud. (Mackay et al. 1998), and as *Aulacoseira* "spore" (Edlund et al. 1995; Stoermer et al. 1995). *A. skvortzowii* is commonly found in its spore morphology in sediments (Edlund et al. 1995, 1996) and modern populations develop in late winter, similar to *A. baicalensis,* under clear ice. It can also be found in abundance in shallow bays and along the thermal bar (Likhoshway et al. 1996). In core BDP 93-2, the absolute abundance distribution of *A. skvortzowii* has two maxima in zones CA-I and IV (Fig. 5e).

Stephanodiscus binderanus (Kütz.) Krieg. and *S. binderanus* var. *baicalensis* Genkal and Popovsk. have been reported from Lake Baikal, and both taxa were present in sediment collections (Genkal and Popovskaya 1990*a*). These taxa differ from each other by the possession of a single central fultoportulae in *S. binderanus* var. *baicalensis*; however, in sediment collections identifications are usually made in girdle view. Hence, the taxa are often indistinguishable, and their species counts were combined. *S. binderanus* is a spring form in Lake Baikal and is present in pelagic collections in the central and southern basins as well as in more nutrient-rich nearshore waters behind the thermal bar (Bradbury et al. 1994; Likhoshway et al. 1996). The distribution of *S. binderanus* in BDP 93-2 is limited to a single absolute abundance peak in the upper seven samples (7–101 cm) of CA-I (Fig. 5f).

Cyclostephanos sp. 1 is an apparently undescribed member of the Baikal phytoplankton community. This taxon has been previously reported as *Cyclostephanos dubius* (Fricke) Round (Bradbury et al. 1994; Stoermer et al. 1995), and it may be a more coarsely ornamented variety of *C. dubius.* The distribution of *Cyclostephanos* sp. 1 in BDP 93-2 was highlighted by a strong peak in CA-I between 69 and 188 cm (Fig. 6a).

Crateriportula inconspicua (I.V.Makarova and Pomazkina) Flower and Håk. was formally described in 1992 (Makarova and Pomazkina 1992), recently transferred to the monotypic genus *Crateriportula* by Flower and Håkansson (1994), and is suspected to be endemic to Lake Baikal and downstream waters. It has been previously reported in Baikal sediment collections as *Stephanodiscus inconspicuus* (Makarova and Pomazkina 1992; Edlund et al. 1995; Stoermer et al. 1995; Mackay et al. 1998) and as *Stephanodiscus* sp. 2 (Khursevich 1989; Bradbury et al. 1994). It was found in surficial sediments from throughout the lake (Stoermer et al. 1995) and has been suggested to be a shallow, warm water, mesotrophic species (Kuzmina and Kobanova 1993; Bradbury et al. 1994). CA-I contained the highest abundance of *C. inconspicua* (Fig. 6b).

Synedra acus Kütz. can be found throughout the year in modern Baikal plankton, but this taxon has its greatest development behind the thermal bar in spring, in shallow bays, and in the Selenga River delta (Bradbury et al. 1994; Likhoshway et al. 1996). It has also been noted to be an increasingly common component of the Baikal phytoplankton
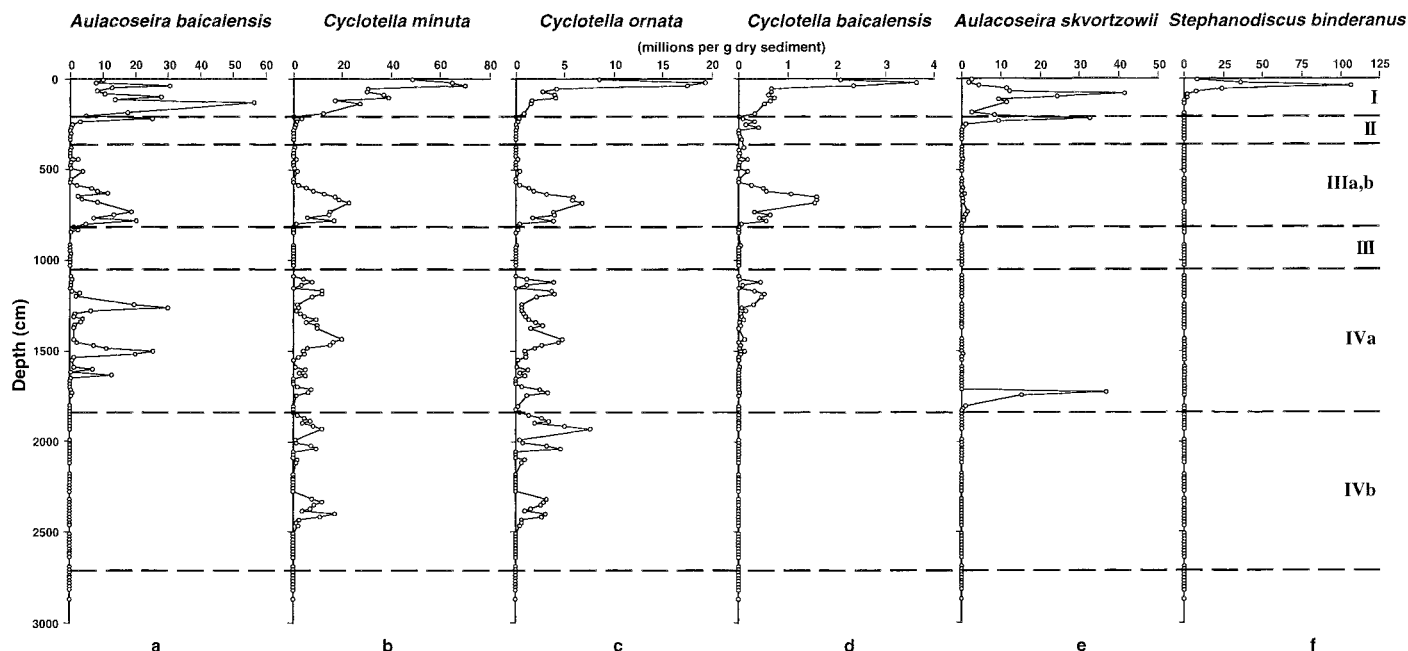
Fig. 5.   Absolute abundance (millions of microfossils per g dry sediment) of siliceous microfossil taxa by sample depth in core BDP 93-2. Discrete sampling depths are represented by open circles in all figures. Lines connecting points are inferred between-sample continuity. Microfossil-depth zones from ordination analysis are shown on right. (a) *A. baicalensis,* (b) *C. minuta,* (c) *C. ornate,* (d) *C. baicalensis,* (e) *A. skvortzowii,* (f) *S. binderanus.*

since the 1950's (Popovskaya 1993). The distribution of *S. acus* is nearly limited to the upper depths of BDP 93-2, with an absolute abundance peak in zone CA-I (Fig. 6c).

Stephanodiscus flabellatus Khursevich and Loginova was described from Lake Baikal sediments and is apparently extinct (Khursevich 1989; Bradbury et al. 1994). Bradbury et

al. (1994) suggested that this taxon may have evolved or become abundant in Lake Baikal during a transition from glacial to interglacial as limnological conditions changed. The distribution of *S. flabellatus* was limited to the upper three zones of BDP 93-2: CA-I, II, and IIIa (Fig. 6d). *S. flabellatus* was conspicuously absent from the top three
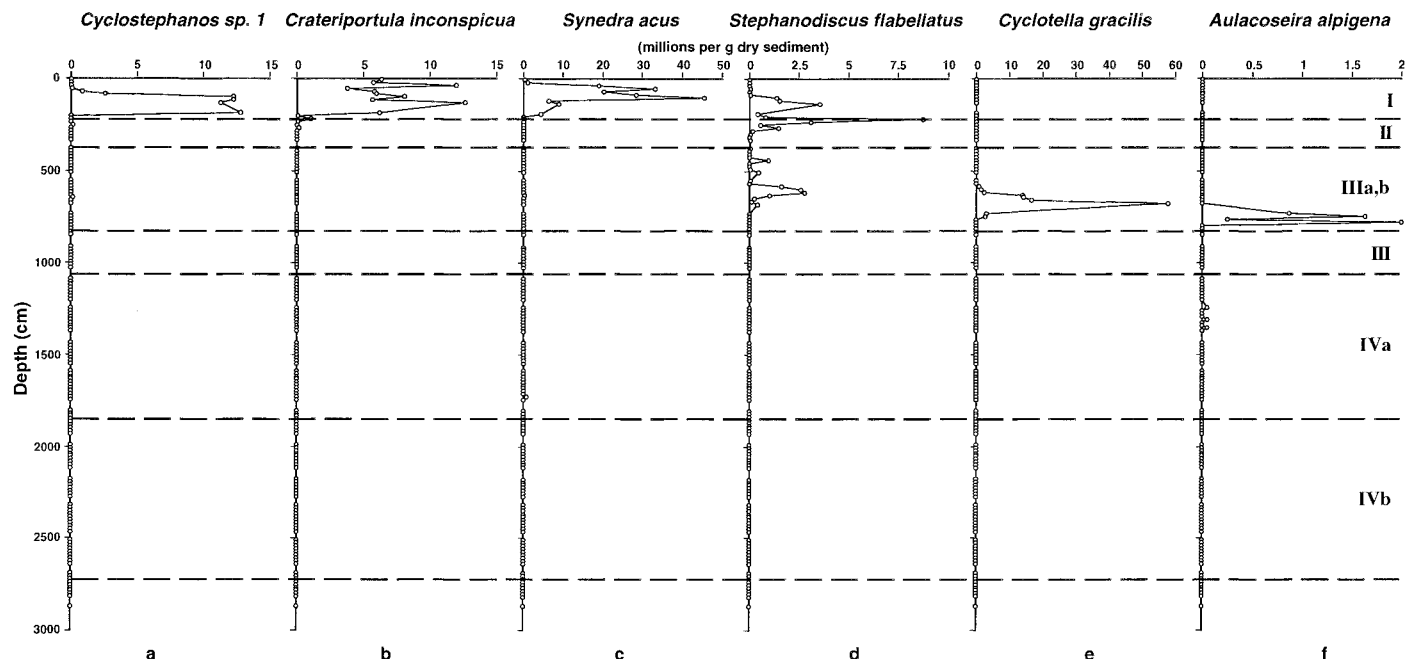

Fig. 6.   as Fig. 5, but (a) *Cyclostephanos* sp. 1, (b) *C. inconspicua,* (c) *S. acus,* (d) *S. flabellatus,* (e) *C. gracilis,* (f) *Aulacoseira* cf. *alpigena.*
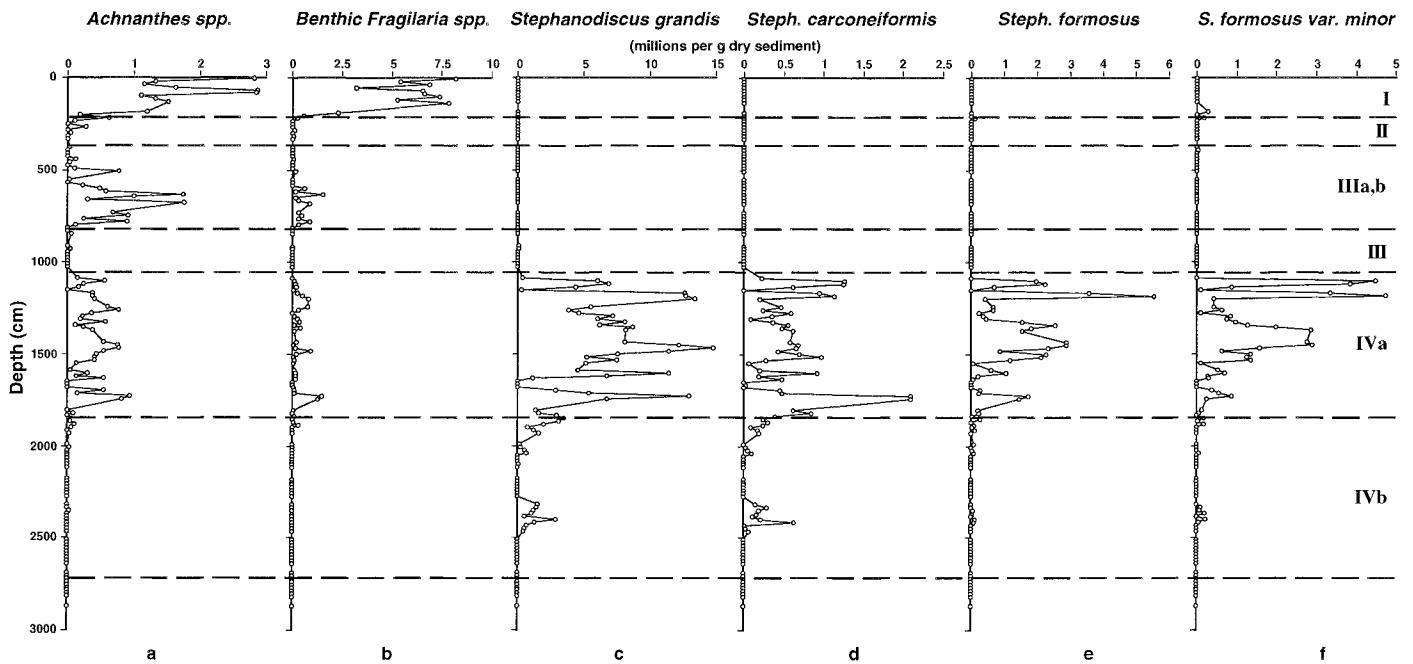
Fig. 7. as Fig. 5, but (a) *Achnanthes* spp., (b) *Fragilaria* spp., (c) *S. grandis,* (d) *S. carconeiformis,* (e) *S. formosus,* (f) *S. formosus* var. *minor.*

depths in the core, placing its extinction date between 2 and 3 kyr B.P.

Cyclotella gracilis Nikiteeva and Likhoshway (1994) was also described from Pleistocene Baikal sediments and is not found in modern Baikal collections. *C. gracilis* has been previously reported as *Cyclotella* sp. (Bradbury et al. 1994) from Baikal sediments and has its distribution in BDP 93-2 limited to zone CA-III (Fig. 6e). A single abundance peak characterizes the downcore profile of *C. gracilis.* Specimens of *C. gracilis* in BDP 93-2 and those illustrated in Bradbury et al. (1994) show a wider size range than noted in this taxon's original description (Nikiteeva and Likhoshway 1994).

Aulacoseira cf. *alpigena* (Grunow) Krammer is a circumboreal taxon commonly recorded from north-temperate and alpine, oligotrophic, soft water lakes (Krammer and Lange-Bertalot 1991). It was found in several depths in BDP 93-2 at low abundance except for a single abundance peak in zone CA-IIIb (Fig. 6f).

Whereas benthic and periphytic diatoms are the most species-rich diatom group in Lake Baikal (Skvortzow 1937; Pomazkina and Votyakova 1993; Stoermer and Edlund 1999), their abundance in sedimentary deposits is usually much lower than planktonic taxa (Fig. 3c,d). Two groups of benthic diatoms, the *Achnanthes* spp. (*sensu lato*) and benthic *Fragilaria* spp. (*sensu lato*), were the only taxa abundant enough to be included in ordination analyses. Both of these groups have recently undergone taxonomic revision at the genus level (e.g., Round et al. 1990); however, they will be treated here in the broader sense. *Achnanthes* species commonly included *Achnanthes striata* Skvortsov and K.I.Mey., *Achnanthes calcar* Cleve, *Achnanthes meyeri* Skvortsov, *Achnanthes minutissima* Kütz., and *Achnanthes oestrupii* (A.Cleve) Hust. The similar ecological role that

Achnanthes spp. have in the periphyton community warrants grouping these taxa. *Fragilaria* species in BDP 93-2 were commonly *Fragilaria pinnata* Ehrenb., *Fragilaria construens* Ehrenb., and *F. construens* var. *minuta* Temp. and Perag. The benthic *Fragilaria* spp. also have a similar ecological habit as periphyton on rock and sand substrates. The appearance of benthic taxa (e.g., *Achnanthes* and *Fragilaria* spp.) in the sediments of Baikal may be explained by occasional entrainment and subsequent maintenance of these populations in the water column resulting from mixing events (Stoermer 1980; Kozhova and Kobanova 1993), by changing lake levels that create relatively more littoral zone, or by diatom production limited to marginal areas during permanent ice cover. *Achnanthes* spp. were distributed in most of the microfossil-bearing sediments in BDP 93-2 and had notable peaks in absolute abundance in zones CA-I, III, and IV (Fig. 7a). Highest peaks in relative abundance of *Achnanthes* spp. were during periods of low microfossil accumulation (with benthics, Fig. 3d). Benthic *Fragilaria* had similar distribution of absolute abundance peaks as *Achnanthes* but were commonly found in higher abundance (Fig. 7b). The highest relative abundances of *Fragilaria* spp. were also observed during periods of low microfossil accumulation (with benthics, Fig. 3d).

Four *Stephanodiscus* taxa are characteristic of zone CA-IV in core BDP 93-2. *S. grandis* Khursevich and Loginova, *Stephanodiscus carconeiformis* Khursevich and Loginova, *Stephanodiscus formosus* Khursevich, and *S. formosus* var. *minor* Khursevich and Loginova were all described from Baikal sediment collections (Khursevich 1989, 1999). *S. formosus* and *S. formosus* var. *minor* were first described and reported from Baikal sediments as *Stephanodiscus bellus* Khurs. and Loginova and *S. bellus* var. *minor* Khurs. and Loginova (Khursevich 1989; Loginova and Khursevich
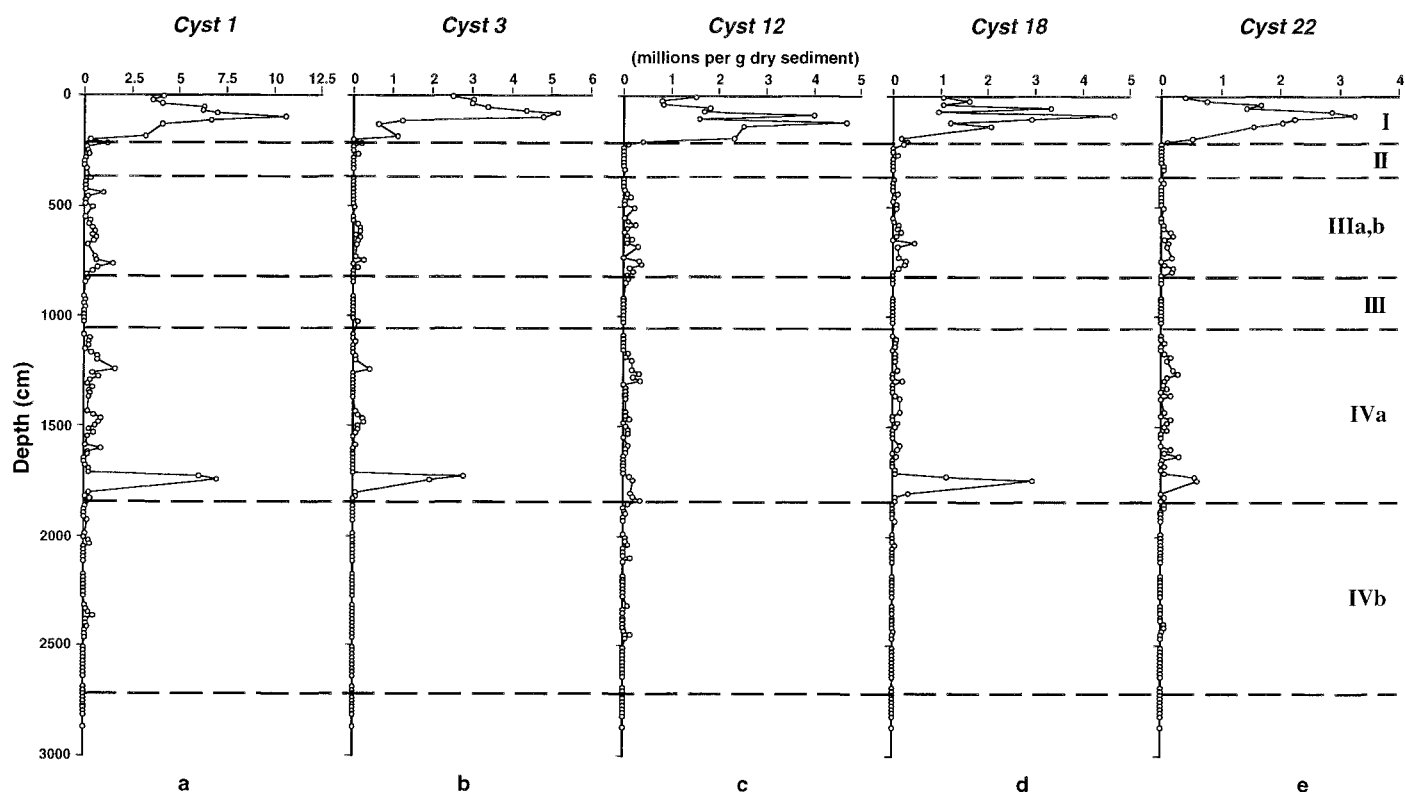
Fig. 8.   as Fig. 5, but (a) chrysophyte cyst 1, (b) chrysophyte cyst 3, (c) chrysophyte cyst 12, (d) chrysophyte cyst 18, (e) chrysophyte cyst 22.

1990; Bradbury et al. 1994; Julius et al. 1997). *S. bellus* Khurs. and Loginova (1990) was shown to be a later homonym of *S. bellus* A. Schmidt. *S. formosus* was proposed as the new name (Khursevich 1999) along with *S. formosus* var. *minor*. All four taxa were apparently limited to biogeographical distributions within the Baikal Rift Zone and are not known from modern collections. Their distributions are essentially limited to zone CA-IV in core BDP 93-2 (Fig. 7c–f), although specimens are found farther downcore (Julius et al. 1997; Edlund 1998). *S. grandis* was the largest and most abundant member of this "archaic" *Stephanodiscus* complex. Because these taxa are all extinct, ecological information can only be inferred from modern *Stephanodiscus* populations. Modern large-celled *Stephanodiscus* (*S. grandis* can reach >150 μm diameter) are found in large-lake systems of varying nutrient regimes but usually require at least moderate silica abundance and regular mixing events for growth (Stoermer and Kreis 1980; Theriot 1992).

Chrysophyte cysts are a conspicuous component of the Baikal sediment assemblage and often had higher relative abundance in levels with low absolute microfossil abundance (e.g., 332–583 cm, Fig. 3e,f). Five cyst morphotypes were found in high abundance, and all were distributed throughout the entire core. Cyst 1 was especially abundant in zone CA-I and near 1,750 cm in zone CA-IV (Fig. 8a). Cyst 3 had similar peak absolute abundance distribution as cyst 1, but was usually found in lower abundance than cyst 1 (Fig. 8b). Cyst 12 had its greatest absolute abundance in zone CA-I (Fig. 8c). Cyst 18 and cyst 22 were distributed similarly to

cysts 1 and 3 with two major absolute peaks in zones CA-I and IV (Fig. 8d,e).

Discussion

Previous studies of the Lake Baikal sediment record have noted changes in both absolute abundance of siliceous microfossils and in the species composition of the preserved microfossil community. Chernyaeva (1970) reported qualitative differences in areal and sediment depth distribution of fossil diatoms in Baikal's north basin. Bezrukova et al. (1991) and Bradbury et al. (1994) independently identified a large increase in accumulation of diatoms in sediments deposited since the start of the Holocene. Qiu et al. (1993) corroborated these observations against similar trends in Holocene biogenic silica accumulation. Bradbury et al. (1994) further characterized Late Pleistocene–Holocene sediment profiles using six dated age zones and two undated zones that were identified by their preserved diatom communities. Similar approaches were used on the longer sediment records collected by the BDP. Core 93-1 was shown to have 12 peaks in microfossil abundance from 0 to 500 kyr B.P. (Julius et al. 1997).

Of the paleoclimate proxies identified in Lake Baikal sediments, biogenic silica has proved most informative. Sedimentary abundance of biogenic silica is highly variable in Baikal cores and inferred to correspond with autochthonous water column productivity, notably diatom production (Ed-

lund et al. 1995; Julius et al. 1997). Periods of increased biogenic silica accumulation are further inferred to represent periods of climatic warming, either interglacial or interstadial (Colman et al. 1995, 1999; Williams et al. 1997). Biogenic silica accumulation in Baikal appears to follow periodicities correlated to Milankovitch orbital parameters, which are believed to be the main external cause of late Pleistocene ice age succession (Colman et al. 1995, 1999; Williams et al. 1997). A recent model relating modern phytoplankton biomass to water temperature lends crucial support to these inferences. Shimaraev et al. (1995) proposed a model based on several decades of Baikal records, in which $\ln B = 0.81 T_w - 1.16$ (no summary statistics provided), where B, the annual biomass of phytoplankton for the photic zone (mg m^{-3}), is positively correlated to T_w, the average surface water temperature from June to October in Southern Baikal.

There is a twofold utility of the siliceous microfossil assemblage record as a paleoclimate indicator in Lake Baikal. The first use follows directly from its relation to biogenic silica accumulation as noted above. High correspondence between microfossil abundance and weight percent biogenic silica is clear (Fig. 2b,c), although correlation ($r^2 = 0.72$) is moderate due to the preponderance of small-sized taxa in Holocene sediments. Microfossil analysis also provides a second, more informative, paleoclimate indicator because it allows reconstruction not only using microfossil abundance as a paleoproductivity proxy, but further reveals the response of the biotic community to historical ecosystem perturbations. Edlund et al. (1995) demonstrated the power of this technique on Baikal sediments by using autecological properties of modern taxa to show that climate fluctuations as short term as the Little Ice Age (ca. A.D. 1630–A.D. 1850 in the Baikal region) produced resolvable differences in pelagic community structure. This approach is appropriate for use in large lake paleo-reconstruction because modern analogs of these systems are unavailable for constructing calibration models (Stoermer et al. 1993).

Whereas changes in microfossil and biogenic silica abundance in the Baikal sediment record have been suggested to represent glacial–interglacial transitions, the correspondence of these two records with the current concept of mid-Siberian highland glacial history has not been established (Julius et al. 1997; Colman et al. 1999). The current Siberian highland glacial record has been generated by palynological records, depositional lithologies, and glacial geomorphological reconstruction (Fig. 9; Arkhipov et al. 1986*b*; Bowen et al. 1986; Velitchko et al. 1989) and indicates that between 0 and 200 kyr B.P., the Siberian highland has undergone three glaciations, three interstadial events, and is currently in an interglacial, the Holocene (Fig. 9). The Baikal microfossil and biogenic silica accumulation records do not completely support the current glacial–interglacial model, as evidenced by poor alignment of productivity peaks and microfossil zones along several well-established glacial–interstadial boundaries (Fig. 9). These observations suggest that an alternative climate response model for Lake Baikal is warranted, based on the microfossil record.

Perhaps the most convincing correspondence of the Baikal microfossil assemblage as a paleoclimate proxy can be seen

when BDP 93-2 microfossil zones are compared to other global paleoclimate proxies. The $\delta^{18}O$ marine isotope record (Imbrie et al. 1984) is perhaps the most commonly used global climate proxy for the Late Pleistocene. The record is based on a transformed ratio of ^{18}O to ^{16}O in fossil marine foraminifera and reflects changes in global ice volume in response to climate shifts (Imbrie et al. 1984). This record has been tuned by several independent research groups (Imbrie et al. 1984; Martinson et al. 1987) by incorporating Milankovitch parameters into the final climate model. This final record has been divided into events or stages that correspond to major and minor inflection points (Fig. 9; Martinson et al. 1987).

The section analyzed in core BDP 93-2 covers from 0 to 200 kyr B.P. and corresponds to marine isotope stages 1–7 (7 only in part). The $\delta^{18}O$ record has been scaled to the corresponding age-depth model of BDP 93-2 in Fig. 9. Alignment of the microfossil zones CA-I to CA-IV to $\delta^{18}O$ stages in this independent climate proxy is exciting. CA-I corresponds precisely to $\delta^{18}O$ stage 1, and CA-II corresponds closely to $\delta^{18}O$ stage 2. $\delta^{18}O$ stage 3 incorporates both CA-IIIa and IIIb, and the remainder of CA-III corresponds to $\delta^{18}O$ stage 4. Sample depths in CA-IV correspond to $\delta^{18}O$ stages 5 and 6. The period between $\delta^{18}O$ stages 5 and 6 was the only transition that was not clearly resolved with correspondence analysis. Upon closer examination of the distribution of sample depths in the ordination, however, it was noted that a differentiation between $\delta^{18}O$ stages 5 and 6 could be found because the younger samples (CA-IVa) were distributed left of center in CA-IV and older samples (CA-IVb) distributed right of center (Fig. 4b). As noted earlier, there were, however, a few sampling depths that were not perfectly grouped within these two zones (Fig. 4b). The alignments of microfossil zones to stages in the $\delta^{18}O$ record indicate that significant changes in the abundance and make-up of the Lake Baikal pelagic community have occurred during periods of rapid climate change.

Zone CA-I (0–11.4 kyr B.P.) resolved the first peak in microfossil abundance and corresponds precisely with the Holocene interglacial (Fig. 9, 0–11.5 kyr B.P.) and $\delta^{18}O$ stage 1 (0–12 kyr B.P.). This period was characterized by high diatom and chrysophyte cyst abundance and the presence of several marker taxa including *S. acus, S. binderanus, Cyclostephanos* sp. 1, and *C. inconspicua.* The Holocene sediment record in Baikal has been the most studied, and historical higher productivity evident in the record is in accordance with the temperature-planktonic productivity model of Shimaraev et al. (1995). Carter and Colman (1994) and Qiu et al. (1993) identified higher biogenic silica accumulation during the Holocene climatic optimum between ca. 4000 and 7000 B.P.; these productivity increases, however, were not clearly evident in the BDP 93-2 record because sampling intervals were too coarse.

Qiu et al. (1993) and Bradbury et al. (1994) noted a hiatus in Holocene biogenic silica deposition from between 8.5 and 11 kyr B.P. that was evident in BDP 93-2 below 204 cm (ca. 11.4 kyr B.P.). The biogenic silica record in this region of core BDP 93-2 was more finely analyzed by Colman et al. (1999) and has been inferred to represent the Younger Dryas cooling (Fig. 9). The Younger Dryas was an early Holocene
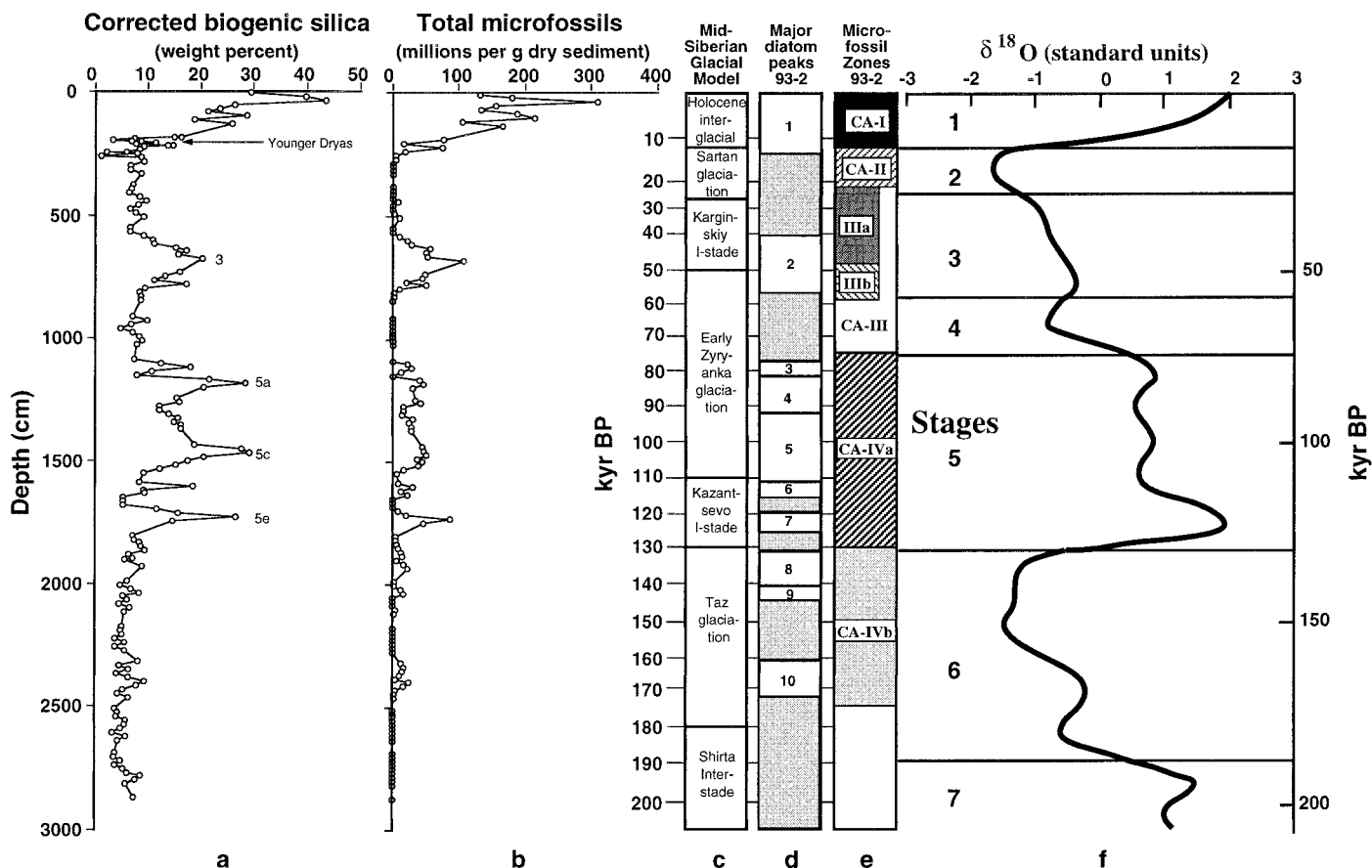
Fig. 9. Summary diagram for core BDP 93-2. (a) Clay-corrected weight percent biogenic silica (Colman et al. 1999), (b) total microfossil abundance (millions of microfossils per g dry sediment). (c–f) Correlative information regarding BDP 93-2 plotted against age model of BDP 93-2 and scaled to core depth. (c) Mid-Siberian highland glacial–interglacial stages. (d) Numbered peak depths of high microfossil abundance in BDP 93-2, shaded areas are periods of low or no total microfossil accumulation. (e) The microfossil zones identified in core BDP 93-2. (f) The marine $\delta^{18}O$ isotope stages (Martinson et al. 1987).

cooling period that occurred between ca. 10–11 kyr B.P. and has been identified in climate reconstructions from locations throughout the world (Peteet 1995). In BDP 93-2, the microfossil accumulation also matches the biogenic silica record with lower absolute microfossil abundance during the Younger Dryas. Detailed reconstruction of the phytoplankton response to this cooling would necessitate finer sampling intervals.

The mid-Siberian Highland glacial model places the Sartan glaciation between 12 and 26 kyr B.P., and BDP 93-2 levels that correspond have noticeably lower microfossil and biogenic silica abundance (Fig. 9). These ages incorporate microfossil zone CA-II (220–332 cm; 12.3–18.7 kyr B.P.); however, the glacial dates extend somewhat deeper into the core to include the upper depths of CA-IIIa (21.3–47.2 kyr B.P.). Microfossil assemblage CA-II corresponds more closely with $\delta^{18}O$ stage 2 (12.0–24.1) than with the glacial model. Several microfossils have absolute and/or relative abundance peaks during the CA-II and in the late Sartan glacial, which suggests greater relative contribution nearshore (benthics, *S. acus, A. skvortzowii*) and/or late-season productivity (*C. minuta*) as expected under a cooler climate (Edlund et al. 1995).

A second major peak in diatom accumulation was re-solved by zones CA-IIIa and CA-IIIb. These two zones together (21.3–56.9 kyr B.P.) correspond closely to $\delta^{18}O$ stage 3 (24.1–58.9 kyr B.P.) and include the mid-Siberian highland model's Karginskiy interstade (26–50 kyr B.P.). Zones CA-IIIa and CA-IIIb began with a large productivity peak that approximates the Karginskiy interstade and ends with a period of relative low microfossil accumulation that may represent the onset of cooling before the Sartan glacial. The microfossil assemblage characterizing the zones CA-IIIa and CA-IIIb includes an initial appearance of *Aulacoseira alpigena* in the Baikal fossil record during CA-IIIb that is followed by the first occurrence of *C. gracilis* in Lake Baikal. The modern Baikal endemic *Aulacoseira* and *Cyclotella* flora were well developed during this time. *S. flabellatus* appears in CA-IIIa during the waning of the Karginskiy interstade. This taxon was apparently well adapted to cooler climatic and limnological conditions as it survived through $\delta^{18}O$ stage 2 but went extinct around the time of the Holocene climatic optimum (Bradbury et al. 1994; Edlund 1998). Competition with the characteristic Holocene floral elements may have also contributed to its demise.

A zone of low microfossil abundance underlies the ca. 58–77 kyr B.P. productivity peak (Fig. 9) and has been identified

within CA-III as a series of ungrouped microfossil sample depths below CA-IIIb. These levels correspond, in part, to the Early Zyryanka glacial, which has been dated from 50 to 110 kyr B.P.; however, correspondence to the glacial model is weak. The correspondence is best to δ^{18}O stage 4 (58.9–73.9 kyr B.P.), and this period must represent a cool climate with low productivity. The preserved microfossil community contains very low abundance but includes the modern *Aulacoseira* and *Cyclotella* flora and the last appearance of *S. grandis,* which suggests that pelagic productivity, albeit low, continued in Lake Baikal even during cold climates. The influence of this cool period on Lake Baikal was very dramatic because many changes in the phytoplankton community were brought about by the short-lived δ^{18}O stage 4, notably the demise and extinction of the ''archaic'' *Stephanodiscus* flora including *S. grandis, S. carconeiformis,* and *S. formosus.*

A remarkable transition in the core occurs below 78 kyr B.P. (ca 10.5 m depth). Below this depth, identified as zone CA-IVa (77.4–129.2 kyr B.P.), the ''archaic'' *Stephanodiscus* flora are found in a series of abundance peaks that continues beyond the lowest microfossil-bearing sediments (2,464 cm) analyzed (Edlund 1998). Zone CA-IVa corresponds with δ^{18}O stage 5 (73.9–129.8) and contains three productivity peaks in microfossil and biogenic silica accumulation that further correspond to δ^{18}O substages 5a, 5c, and 5e (Fig. 9). Lengthy interstadial periods have been shown to have subperiods of warming and cooling that might produce such peaks (Arkhipov 1984). The Kazantsevo interstade is dated from between 110 and 130 kyr B.P.; however, from the Baikal record, the warming associated with δ^{18}O stage 5 and zone CA-IVa is inferred to have lasted from ca. 130 to 78 kyr B.P. The end of this interstade changed conditions in Lake Baikal sufficiently to cause the extirpation of the ''archaic'' *Stephanodiscus.* The establishment of *A. baicalensis* and the apparent evolution of *C. baicalensis* during this final period of *Stephanodiscus* dominance (Julius et al. 1997; Edlund 1998) in Lake Baikal may have contributed to the extirpation of these flora. Conditions in Lake Baikal during δ^{18}O stage 5 must have approached near-modern temperature regimes as biogenic silica accumulation rates approached Holocene levels (Shimaraev et al. 1995; Colman et al. 1999). The predominance of large-celled taxa during this time dampens the microfossil abundance profiles somewhat compared to Holocene times. From our knowledge of autecological distribution of modern large-celled *Stephanodiscus* (Stoermer and Kreis 1980; Theriot 1992) and the modern Baikalian endemics (*A. baicalensis, Cyclotella* spp.), these characteristic floral elements suggest that Lake Baikal was experiencing deep-mixing events and clear, winter ice cover (Edlund et al. 1995, 1996).

Below the earliest peak in the δ^{18}O stage 5 and zone CA-IVa (ca. 130 kyr B.P.), microfossil and biogenic silica abundances were lower. There are several microfossil peaks below this depth; however, their magnitudes are much smaller than during δ^{18}O stage 5. These lower microfossil-bearing reaches were resolved as CA-IVb (130.3–172.2 kyr B.P.) and, including the barren zone below (2,510–2,870 cm), correspond to δ^{18}O stage 6 (129.8–189.6 kyr B.P.), a lengthy stage of fluctuating cool climates that allowed some periods

of moderate productivity. During the short periods of productivity within this cool stage, the flora was characterized by the ''archaic'' *Stephanodiscus, C. minuta,* and *C. ornata.* δ^{18}O stage 6 and CA-IVb correspond closely to the Taz glacial, the last glacial period in the Bakhta glacial epoch that is dated in the mid-Siberian glacial model at 130–180 kyr B.P. Samples immediately below 2,870 cm are microfossilladen in BDP 93-2 and inferred to represent the Shirta interstade and the warm δ^{18}O stage 7.

Some taxa identified in core BDP 93-2 may be important as stratigraphic indicators due to their presence in few sampling depths or by well-defined distribution boundaries. Distributions of indicator species are efficient means of identifying stratigraphic facies, continuity between samples (Bradbury et al. 1994), and periods of severe ecological perturbation. Indicator species could be used in future efforts on the sedimentary record from Lake Baikal and other basins in the Baikal Rift Zone.

The extirpation of the ''archaic'' *Stephanodiscus* flora is one such stratigraphic indicator. These flora are characteristic of zone CA-IVa and CA-IVb and is limited in sedimentary distribution to periods earlier than 77 kyr B.P. (Fig. 7c–f). The ''archaic'' *Stephanodiscus* flora was made up of four taxa, *S. grandis, S. carconeiformis, S. formosus,* and *S. formosus* var. *minor,* which were limited to existence during the Pleistocene. They apparently went extinct in Southern Baikal ca. 77 kyr B.P., at the onset of δ^{18}O stage 4, during initiation of cooler climate conditions.

C. gracilis was found in abundance in Baikal sediments between 40 and 53 kyr B.P. Its distribution was limited to middle and late Karginskiy interstade sediments in zones CA-IIIa,b, and its extinction appears to coincide with a climatic shift into the Sartan glaciation and δ^{18}O stage 2. *C. gracilis* was also used as a stratigraphic indicator by Bradbury et al. (zone VII, 1994); however, they were unable to place it in association with a dated sediment profile.

Aulacoseira cf. *alpigena* was found in abundance only in sediments dated from ca. 50 to 55 kyr B.P. in CA-IIIb. This microfossil zone apparently represents the initial increase in phytoplankton production at the onset of δ^{18}O stage 3 (Fig. 9). The appearance and subsequent survival of this taxon in Lake Baikal is odd among the other taxa encountered in the core, and it suggests that cosmopolitan taxa have become established for short periods of time in the Baikal phytoplankton following climate shifts. *A. alpigena* is more commonly noted in north-temperate, soft water lake systems (Krammer and Lange-Bertalot 1991).

A final stratigraphic indicator noted in BDP 93-2 was the assemblage deposited during the Holocene. In addition to high abundances of the more showy endemic Baikalian diatoms (*A. baicalensis, C. minuta, C. ornata, C. baicalensis*), several taxa can be used as stratigraphic indicators of Holocene deposition. *S. acus, C. inconspicua, S. binderanus,* and *Cyclostephanos* sp. 1 were deposited in high abundance only during Holocene periods identified in this study as CA-I. *S. binderanus* is especially characteristic of Holocene sediments because it was limited to sediments deposited between 0 and 6 kyr B.P.

The discordance between the glacial and interglacial cycles and the Baikal microfossil record probably resulted from

the fact that major ice sheets involved in Pleistocene glaciations were limited in their expansion to more northerly continental Siberian latitudes (Arkhipov et al. 1986*b*). In turn, this may have produced more moderate climate impacts in and around Lake Baikal, especially during glacial periods, and the temporal incongruities that are evidenced in the BDP 93-2 record. Evidence of marginal montane glaciers penetrating into Lake Baikal at the end of the Pleistocene has been found in Baikal's north basin (Back et al. 1998, 1999) and indicates that cooler climates were impacting the system with colder air and water temperatures and increased sediment inputs but without creating permanent ice cover. This would have resulted in lower annual productivity and less autochthonous sedimentation.

The microfossil record is more strongly temporally aligned with the $\delta^{18}O$ marine record vs. glacial/interglacial cycles. Each microfossil assemblage zone identified with correspondence analysis approximates a stage in the $\delta^{18}O$ climate model. Even minor events in the marine isotope record are correlated to peaks in microfossil and biogenic silica accumulation in BDP 93-2. Colman et al. (1999) noted that $\delta^{18}O$ stages 3, 5a, 5c, and 5e were represented by silica peaks at 680, 1,210, 1,460, and 1,730 cm, respectively (Fig. 9). Karabanov et al. (1998) suggested that $\delta^{18}O$ stage 5d was recorded in BDP-93 as an abrupt glaciation that followed the Kazantsevo interstade from evidence of low biogenic silica accumulation at 1,550–1,650 cm. The minor $\delta^{18}O$ stages are corroborated in the microfossil record by identical peaks and valleys in absolute microfossil abundance.

The alignment between microfossil zones and the $\delta^{18}O$ marine isotope record indicates that primary producer communities in Lake Baikal have responded to orbitally induced climatic variation on a time frame more similar to the tropical and subpolar oceans than to the mid-Siberian Highlands. Further evidence of a Baikalian-oceanic climate connection is seen in decadal-level signals of North Atlantic Oscillation climate variability in modern Baikal ice-out dates (Livingstone 1999). The response of Baikal's primary producer community to climatic variation has been twofold. First, colder climate stages are characterized by lower productivity and microfossil accumulation, and, more strikingly, each climate stage during the history of Lake Baikal had a unique microfossil assemblage. Climate changes impact Lake Baikal with biotic upheaval through extinction and origination of taxa. The preserved diatom assemblage provides one of the few records of community-level response to climatic perturbations and a proxy for the biotic response potential of the highly endemic Baikal ecosystem to future climate change.

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