Gas buildup in Lake Nyos, Cameroon: The recharge process and its consequences

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Abstract—The gases dissolved in Lake Nyos, Cameroon, were quantified recently (December 1989 and September 1990) by two independent techniques: *in-situ* measurements using a newly designed probe and laboratory analyses of samples collected in pre-evacuated stainless steel cylinders. The highest concentrations of CO_2 and CH_4 were 0.30 mol/kg and 1.7 mmol/kg, respectively, measured in cylinders collected 1 m above lake bottom. Probe measurements of *in-situ* gas pressure at three different stations showed that horizontal variations in total dissolved gas were negligible. Total dissolved-gas pressure near the lake bottom is 1.06 MPa (10.5 atm), 50% as high as the hydrostatic pressure of 2.1 MPa (21 atm). Comparing the CO_2 profile constructed from the 1990 data to one obtained in May 1987 shows that CO_2 concentrations have increased at depths below 150 m. Based on these profiles, the average rate of CO_2 input to bottom waters was 2.6×10^8 mol/a. Increased deep-water temperatures require an average heat flow of 0.32 MW into the hypolimnion over the same time period. The transport rates of CO_2, heat, and major ions into the hypolimnion suggest that a low-temperature reservoir of free CO_2 exists a short distance below lake bottom and that convective cycling of lake water through the sediments is involved in transporting the CO_2 into the lake from the underlying diatreme. Increased CH_4 concentrations at all depths below the oxycline and a high ^{14}C content (41% modern) in the CH_4 4 m above lake bottom show that much of the CH_4 is biologically produced within the lake. The CH_4 production rate may vary with time, but if the CO_2 recharge rate remains constant, CO_2 saturation of the entire hypolimnion below 50 m depth would require ~140 a, given present-day concentrations.

INTRODUCTION

THE DEVASTATING release of CO_2 from Lake Nyos (Fig. 1) on 21 August 1986 resulted in the deaths of 1700 people. Overshadowing a smaller but similar event at Lake Monoun on 15 August 1984 (SIGURDS-SON *et al.*, 1987), the Lake Nyos disaster attracted international attention, and the results of a number of scientific studies have recently been compiled (LE GUERN and SIGVALDASON, 1989, 1990). Debate focused on whether the disaster was caused by a volcanic (phreatic) eruption from beneath the maar lake (TAZIEFF, 1989; BARBERI *et al.*, 1989), a limnological disturbance releasing magmatic CO_2 previously stored within the lake (TUTTLE *et al.*, 1987; TIETZE,

1987: KLING *et al.*, 1987, 1989; KUSAKABE *et al.*, 1989; GIGGENBACH, 1990), or a combination of both processes (FREETH and KAY, 1987).

One test of the limnological hypothesis involves the recharge of CO_2 into the lake. While a long-term flux of CO_2 into bottom waters is compatible with the volcanic eruption theory, it is critical to the limnological explanation. If the gas was slowly building up within the lake at the time of the gas burst, the influx of gas from below should continue after the burst. To test the validity of the limnological hypothesis and to gauge the likelihood of a recurrent gas burst from Lake Nyos, one of the most important parameters to monitor is dissolved gas pressure in the hypolimnion. Partly because the gas pressures are much higher
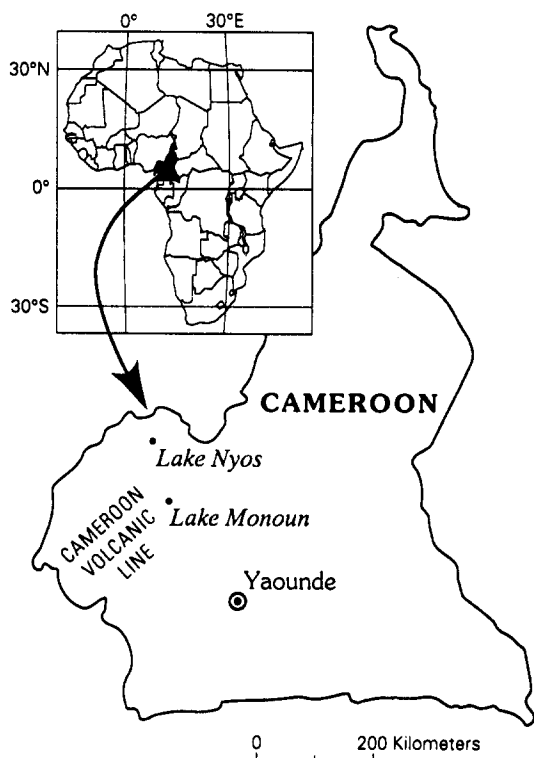
FIG. 1. Map of Cameroon showing lake locations.

than those usually encountered in lakes, accurate measurement of the dissolved gases has proven very difficult.

Previous gas studies

The earliest gas samples collected in September 1986, two weeks after the burst, showed that CO_2 comprised >99% of the gas dissolved in the hypolimnion (TUTTLE *et al.*, 1987), but gas-to-water ratios were not measured. The amount of gas exsolved from subsurface water samples was first measured in October 1986 (KUSAKABE *et al.*, 1989), and dissolved gas concentrations in pressurized samples were first measured in November 1986 (TIETZE, 1987). These studies demonstrated that the dissolved CO_2 profile was similar to the profiles of temperature and total dissolved solids, which increase as a function of depth. To test for CO_2 recharge, samples were collected in pre-evacuated stainless steel cylinders in January, March and May 1987 (KLING *et al.*, 1989). At most depths, the CO_2 profiles were similar to the one reported by TIETZE (1987). However, cylinders filled near lake bottom indicated fairly large (up to 50%) horizontal variations in CO_2 concentrations. This was puzzling in view of the smaller horizontal variations in temperature and ionic strength. At that time, it was not clear whether these CO_2 variations existed in the lake or were caused by gas leakage from some of the cylinders. Before this question was

resolved by later sampling (as discussed below), two cylinder samples were collected on a brief trip in June 1988.

NOJIRI *et al.* (1990) were successful in showing that CO_2 concentrations in December 1988 at 190 and 200 m depths had increased from previous values, although the 200 m measurement was imprecise due to the large volume of gas exsolved. Extrapolating the CO_2–TDS relation from 190 m depth to lake bottom at 210 m, they calculated a CO_2 flux into Lake Nyos of 5.15×10^8 mol/a, and later revised this to 1×10^9 mol/a (NOJIRI *et al.*, 1993). A higher CO_2 flux (2.0×10^9 mol/a) was derived from the December 1988 measurements of temperature and dissolved He by SANO *et al.* (1990). These important studies documented CO_2 recharge. However, the rate calculations involved an assumed relation between CO_2 and another quantity actually measured (TDS or He), and the possibility of horizontal variations in gas concentrations remained untested. Reported here are the results of two new gas surveys of Lake Nyos, carried out in December 1989 and September 1990, using numerous stainless steel cylinders and a new probe designed specifically for real-time measurements of dissolved gas pressure in the hypolimnion.

METHODS OF SAMPLING AND ANALYSIS

Three sampling stations were established on Lake Nyos in December 1989, one of which was reoccupied in September 1990. All stations were located in the central basin, an oval crater with a 0.4 km^2 flat bottom at 210 m depth (Fig. 2). To assure verticality of the sampling line, three fluke-type anchors were set in a triangular pattern around each station. Temperature profiles were measured using a four-conductor thermistor ohm-meter (SASS *et al.*, 1981) calibrated just prior to the field trip and rechecked in the laboratory against a recently certified NBS-traceable thermometer at the conclusion of the trip. The error was within the 0.005°C readability of the thermometer. Samples of water were collected at various depths using a Niskin sampler equipped with a vent valve that allowed exsolving gas to escape during retrieval. Samples were handled as on previous occasions (KLING *et al.*, 1989), i.e. with on-site filtration and alkalinity being titrated before there was any visible precipitation of solids.

Cylinder procedures

The stainless steel cylinders were fitted with two ball valves and one check valve (Fig. 3a). The top ball valve was attached to a trigger on the sampling line and could be opened at depth by a sliding messenger. With this valve open, hydrostatic pressure forces the check valve open (nominal cracking pressure is small, 7 kPa) and fluid enters the cylinder. When internal and external pressures are virtually equal, the check valve closes to trap the sample. Upon cylinder retrieval, the top ball valve was manually closed as a backup seal.

The concentrations of dissolved gases in the cylinders were determined in the laboratory after attaching a headspace cylinder to each sample cylinder (Fig. 3b). Sample fluids were equilibrated with the initially evacuated headspace at 25.0°C to partition the gases between solution and

headspace. The partial pressures of the gases in the head-space were determined by gas chromatography using a thermal conductivity detector. Cylinder volumes are known and the ideal gas law was used to calculate the amount of each gas extracted, except for CO_2 where the non-ideality correction of WEISS (1974) was made. The results were corrected for the gas remaining in solution using the solubility data of WEISS (1974) for CO_2 and WILHELM et al. (1977) for other species. Following gas chromatographic analysis, CO_2 and CH_4 were collected by repeated extraction of the solution. Concentrations of CO_2 previously calculated by gas chromatography were then verified by manometric measurement before the CO_2 was analyzed isotopically. Extracted CH_4 was converted to CO_2 and H_2 prior to isotope analysis. After gas extraction, the water was drained from the cylinders and a small amount analyzed for the C-isotope composition of the HCO_3 and any unextracted CO_2. These results were used to make small corrections to the isotope composition of the extracted CO_2 to generate $\delta^{13}C$ values for the dissolved inorganic carbon (DIC). The remainder of the water was occasionally used for major-ion analysis. The analytical procedures for the cylinder samples were generally similar to those used previously on low-pressure samples described by EVANS et al. (1988).

All cylinders were tested for leaks during evacuation prior to fieldwork. Ten cylinders were evacuated and stored for six months to test long-term leak rates; only one had an air leak rate $>20 \mu mol/month$. Two cylinders were further tested by sampling a pressurized test solution of CO_2 charged, air-saturated water (ASW), from a large bomb. The CO_2 pressure in the bomb was just over 0.4 MPa, in the range of subsurface samples from Lake Nyos. The CO_2 concentrations measured in both cylinders were within 2% of the correct value and measured $\delta^{13}C$ values on the DIC were <0.1‰ in error. Measured amounts of the atmospheric gases, N_2 and Ar, deviated as much as 10% from the correct ASW values, indicating that there were minor

FIG. 3. (a) Sample cylinder deployed, (b) Sample prepared for analysis by attaching headspace cylinder.

problems associated with sampling these highly insoluble gases in the presence of large excess of CO_2. Presumably, this applied also to CH_4 in the samples from Lake Nyos. However, there was no evidence from laboratory testing that cylinder failure could explain the magnitude of the horizontal variations in dissolved gas observed in the 1987 studies.

Probe procedures

Plastic tubing has long been used as a semi-permeable membrane to monitor dissolved gases (ENNS et al., 1965); molecules of each gas species diffuse through the membrane until the fugacities inside the tubing and in the surrounding fluid are equal. The gas probe (Fig. 4) was constructed using medical-grade silicone tubing 1.47 mm I.D. with a 0.24 mm wall thickness. The tubing was cut into 70-cm lengths and 15 of these were attached by slip fit to short lengths of 1/16″ O.D. stainless steel capillary tubing. Braided 18-gauge steel wire was fed through the entire length of the silicone tubing and into the stainless steel capillary to prevent collapse of the silicone tubing under high hydrostatic pressure. The free ends of the silicone tubes were then plugged. The 15
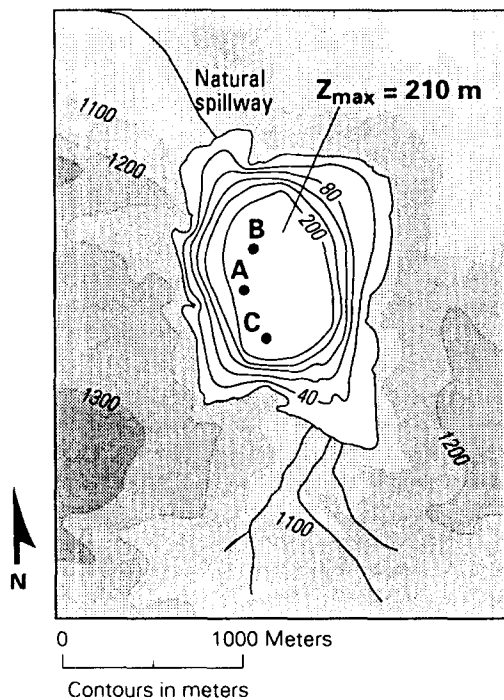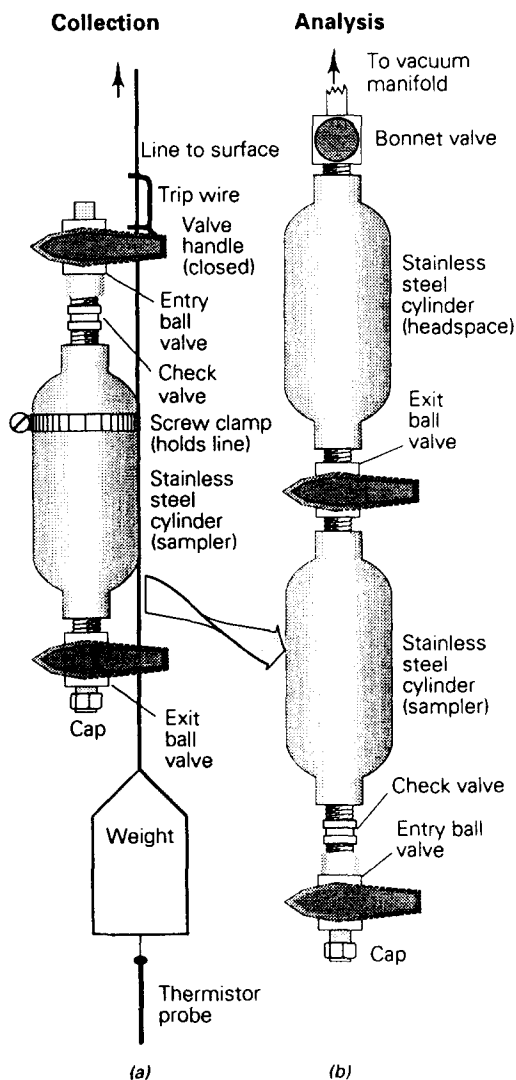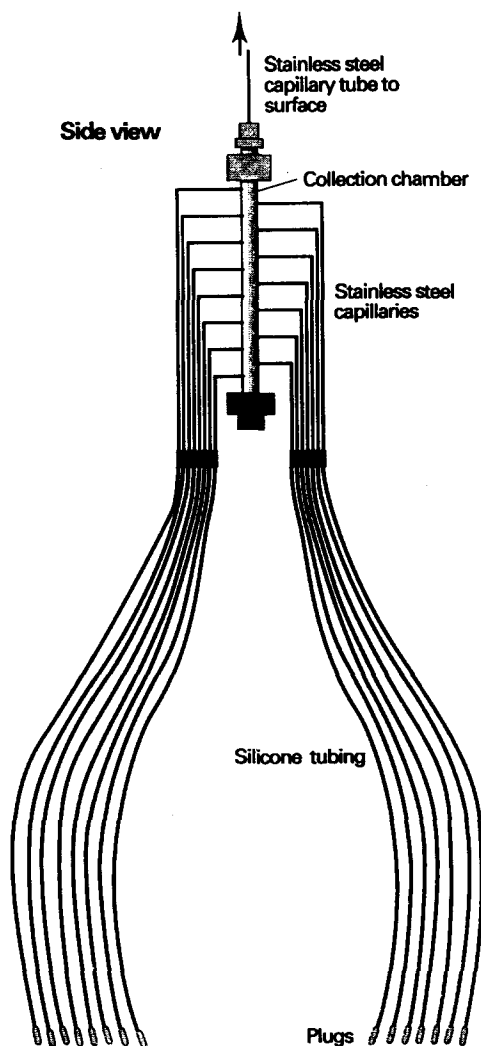
FIG. 2. Bathymetric map of Lake Nyos modified from TUTTLE et al. (1987) showing sampling stations A, B, and C on the 1989 and 1990 expeditions. Depths are from lowest point on the natural spillway.

FIG. 4. Gas-pressure probe.

RESULTS AND DISCUSSION

In Fig. 5, the 1989 and 1990 temperature profiles are compared to those from January and May 1987. Water chemistry data from selected depths are reported in Table 1. Figure 6 shows the profile of alkalinity (as HCO_3) from all sampling dates and depths. Profiles of the other major ions are all similar to that of HCO_3, which makes up >98% of the anions.

The August 1986 gas burst coincided with a partial mixing between cool, dilute surface water and warm, mineralized bottom water affecting the temperature and chemical structure within the lake (TUTTLE *et al.*, 1987; KLING *et al.*, 1987). KLING *et al.* (1989) have discussed the changes that occurred in the ensuing nine months (to May 1987). Rainfall, runoff and surface mixing caused dilution and cooling of the subsurface water to a depth of about 20–30 m, while recharge from warm, mineralized fluid at lake bottom caused temperature and ionic strength to increase in the deepest layers. Between 30 and 175 m depth, temperature and water chemistry did not show significant changes. It was therefore assumed that transport of heat and dissolved chemical species through this zone by fluid convection was insignificant; and ignoring other heat losses (conductive, radiative, etc.), a heat flow of 1600 mW/m^2 (0.93 MW) was required to explain the increased heat content below 175 m depth.

The new temperature profiles show that during the subsequent forty months (May 1987 to September 1990), temperature at lake bottom continued to rise (Fig. 5), but the average rate of increase has slowed from 0.05°C/month in early 1987 to 0.02°C/month between May 1987 and December 1989 and to <0.01°C/month between December 1989 and September 1990. Both new profiles show temperature increases much farther above lake bottom than in May 1987 with increases of >0.05°C up to ~140 m depth. Subsurface cooling has extended to greater depths (to ~75 m). Between 75 and 140 m depth, temperature differences between all profiles are ≤0.05°C. Although it was difficult to verify the accuracy of the thermistor to better than 0.05°C during field use, there have been minor changes in profile shape in this region that cannot be attributed totally to shifts in equipment calibration. The effects of heat transport within this zone are therefore discernible even if very small compared to the effects above and below this zone. The small temperature increase seen in both the new profiles up to ~100 m depth was thus included in the heat flow calculations. The increase in hypolimnetic heat content below 100 m depth was 8.1 × 10^9 kcal between May 1987 and September 1990 and requires a heat flow of 0.32 MW, approximately one-third as much as before May 1987. NOJIRI *et al.* (1993) also report a lower heat flow of 0.43 MW for the 25-month period ending December 1988. Apparently, the heat flow at lake bottom has decreased

stainless steel capillaries were brazed to a 2 cm^3 collection chamber, and the chamber in turn was connected to a 215-m length of 1/16″ O.D. stainless steel capillary tubing that was coiled on a reel. When the probe is immersed in water, gas diffuses through the probe and flows through the capillary tubing to a pressure gauge. The long capillary naturally causes gauge response to be slow. Laboratory testing with the probe sealed in a pressure bomb containing aqueous solutions of CO_2 (pCO_2 = 0.3–1.0 MPa) showed that several hours were required for gauge pressure to equal dissolved gas pressure to within the accuracy of the gas (±10 kPa). However, gauge response curves were of such uniform shape that final pressures could be extrapolated after only 1 h with a total error of ±15 kPa. When deployed, gauge pressures at the lake surface were less than gas pressure at depth because of the weight of the gas in the long column. Gauge pressures were corrected to *in-situ* pressures using:

$$P_0 = (P)\, e^{-Mgz/RT} \qquad (1)$$

where P_0 is *in-situ* pressure, P is gauge pressure, M is the molecular weight of the gas, g is the gravitational acceleration, z is the measurement depth, R is the gas constant and T is absolute temperature. The gas is nearly pure CO_2 and M was assumed to be 44 g/mol. The calculated pressure correction is 3.7% at lake bottom.
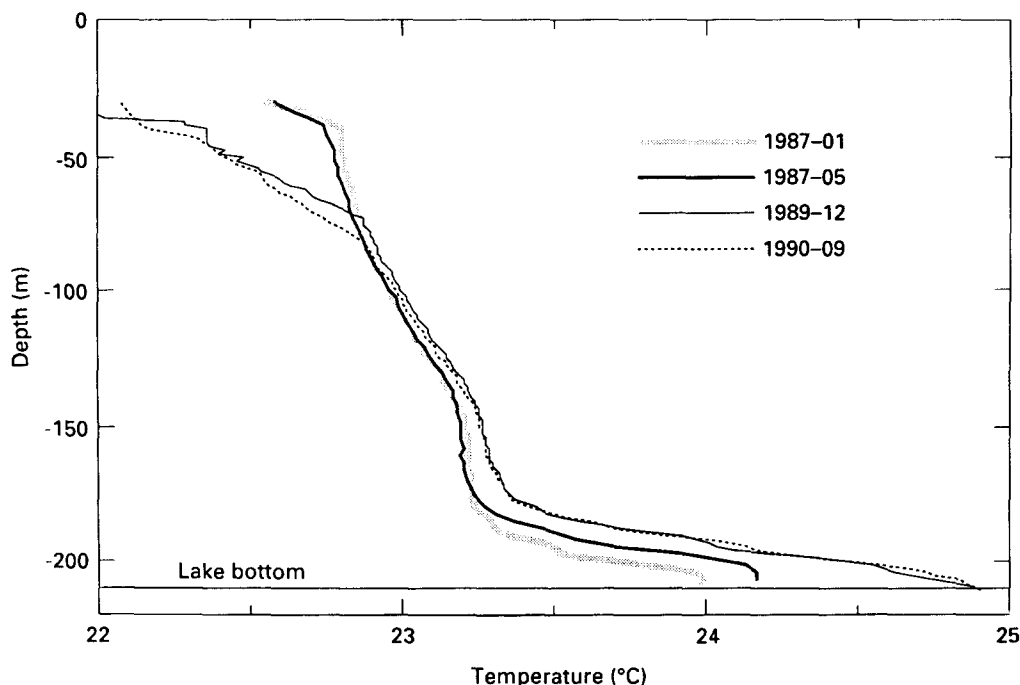
FIG. 5. Temperature vs depth. The large diurnal and seasonal variations in the surface layers (<30 m depth) are not shown, but all data are available on floppy disk from the authors. The 1987 profiles are reported in KLING et al. (1989).

markedly since the period immediately following the gas burst (note that heat losses are ignored in all the heat flow calculations).

The water chemistry data (Table 1, Fig. 6) show continued dilution of surface water. Subsurface dilution due to mixing has gradually reached to ~50 m depth (Fig. 6), not quite as deep as subsurface cooling (Fig. 5). At mid-depths (i.e. 99 m) where temperature changes have been small, the water chemistry also has changed little (Table 1, Fig. 6). Apparently, chemical transport through this zone is still extremely limited. In bottom waters (i.e. 198 m depth), ionic strength continued to increase after May 1987, but values have levelled off showing only small changes after June 1988. Thus the water chemistry and temperature profiles have evolved in similar, but not identical, patterns. Small differences in their evolutions are expected because of the great disparity in the diffusion coefficients of heat and chemical species as discussed by TIETZE (1987) and SABROUX et al. (1990). Thermal perturbations will travel through the water column much faster than chemical ones. The water chemistry data have been discussed in greater detail and used to develop a hydrologic budget for the lake by TUTTLE et al. (1992).

The gas probe results are shown in Table 2. In one case on each expedition, the probe was left stationary long enough for the gauge to reach a final, stable pressure. The gauge response curves (Fig. 7) match closely the shape of curves generated in laboratory tests. On the 1989 expedition, horizontal variations in gas pressure at three depths, 168, 195 and 202 m,

were found to be no greater than measurement uncertainty (Table 2), showing that total dissolved gas, like temperature and major-ion chemistry, is a function only of depth, not horizontal position. Therefore, only one station (C) was reoccupied in 1990 to test for gas pressure changes. The 1990 data show distinct gas pressure increases below 168 m depth during the 10-month interval (Table 2), reaching 50% of hydrostatic pressure at 206 m depth. In one instance, the probe was heavily weighted and allowed to sink into the sediment to test for a sharp pressure change at the sediment–water interface. The pressure measured was slightly less than in the overlying water, although it is not certain that the probe can perform accurately in the muddy sediment.

Cylinder data from selected depths are shown in Table 3. It is likely that most or all of the O_2 detected is from air leakage prior to sample collecuon; the high ferrous iron content of hypolimnetic water indicates strongly reducing conditions. Some of the variations in Ar and N_2 are also caused by air leakage, but even in the most contaminated samples, the amounts of air-derived CO_2 and CH_4 are trivial.

The 1989 sampling provided the first opportunity to test the in-situ performance of the cylinders. Because the probe showed negligible horizontal variations in gas pressure, all cylinders filled at a given depth should contain similar gas concentrations. For two cylinders filled at 137 m depth, concentrations of CH_4 and CO_2 agree within a few percent. The same is true for a pair of cylinders from 171 m depth even

Table 1. Major-ion concentrations* (in mg/kg) at selected depths.

Depth[†] (m)	DATE[‡]	Na	K	Mg	Ca	Fe	HCO$_3$	I§ × 10^3
0	1986-09	6.4	3	16	14	1	165	3.45
	1987-01	6.8	2	18	13	<0.1	161	3.62
	1987-05	7.3	3	19	16	<0.1	163	3.90
	1988-06	6.3	1.8	9.5	9.2	0.2	92	2.16
	1989-12 (2)	3.4	1.4	4.8	4.8	0.2	54	1.18
	1990-09	3	1	4	5	<0.1	49	1.06
16	1987-01	7	3	21	17	3.5	204	4.56
	1987-05	8	3	22	18	0.4	192	4.51
	1988-06	6	1.9	12	11	<0.1	107	2.57
	1989-12	3.8	1.4	5.3	5.3	<0.1	54	1.24
	1990-09	3	1	5	6	<0.1	57	1.26
99	1987-03	13	4	43	34	63	558	12.40
	1987-05	14	4	49	38	58	554	12.90
	1988-06	15	4	45	37	60	531	12.43
	1989-12(3)	14	4	45	35	59	536	12.31
	1990-09	14	4	44	35	60	532	12.23
171	1987-05	18	5.1	59	45	76	686	15.90
	1989-12	17	5.1	55	45	83	694	15.87
	1990-09	16	5	54	42	76	693	15.35
198	1986-09	19	7	57	46	71	725	15.97
	1987-01(4)	20	6	68	47	81	767	17.64
	1987-03(2)	22	7	80	62	106	962	21.93
	1987-05(4)	24	7	85	65	96	963	22.18
	1988-06	25	7	90	71	110	1060	24.21
	1989-12(2)	24	6.2	88	66	121	1053	24.10
	1990-09	23	7	87	66	120	1039	23.86
206	1989-12(3)	28	8	107	79	151	1253	29.14
	1990-09	27	8	100	77	140	1224	27.81

* Analytical uncertainties are ±5% for concentrations reported to two or more significant figures and are otherwise ±1 mg/kg. The Fe is present in divalent form and the HCO$_3$ values represent total alkalinity. Concentrations of Cl, F and SO$_4$ are ≤1 mg/kg at all depths.

† Depths are from lowest point on spillway and are ±1 m on May 1987, December 1989 and September 1990 samples when collection raft was anchored. Greater uncertainties exist on other dates due to drifting.

‡ Analyses from September 1986 are reported in Tuttle *et al.* (1987). Analyses from January 1987, March 1987 and May 1987 are reported in Kling *et al.* (1989). On dates when multiple samples were collected, the number of samples is given in parentheses and the reported analyses are averages.

§ Ionic strengths (I) are calculated from the reported ion concentrations.

though one of the cylinders was heavily air-contaminated. However, the two cylinders filled at 198 m depth differ considerably in CO$_2$ and CH$_4$ concentrations, 18 and 44%, respectively. Three cylinders filled at 206 m depth (including a pair filled simultaneously) vary by 8% in CO$_2$ and 53% in CH$_4$ concentrations. Thus, gas loss from some of the cylinders was worse than laboratory testing had suggested.

The comparison between probe and cylinder data is made more rigorous by using the gas concentrations measured in the cylinders (Table 3) to calculate the *in-situ* Henry's law gas pressure for the sampled fluids. These calculated pressures are shown in Fig. 8a, where they are compared to a smooth curve drawn through the probe measurements (Table 2). With small extrapolations to the probe curve, a comparison is possible for nine cylinder samples. For seven of these, the calculated pressures deviate <5% from the probe curve, showing that cylinder perform-

ance is reasonably good in most cases. The other two cylinders have allowed significant gas loss. Subsequent laboratory testing of these two cylinders indicated that they could hold high gas pressures for many months without leakage. It was thus postulated that most gas loss problems must occur during cylinder filling, probably incomplete opening of the entry ball valve causing gas bubbles to form between it and the check valve (Fig. 3a). These bubbles would be free to escape before the cylinders could be retrieved.

For the 1990 sampling, two changes were made to improve cylinder triggering. A better trip mechanism was installed to cause fast and complete opening of the entry ball valve. Also, cylinders were triggered in an inverted position so that any bubbles formed in the ball valve would be driven by buoyancy through the check valve into the cylinder chamber. Probe and cylinder gas pressures from station 90C are shown in Fig. 8b. Deviations are <5% for all seven cylinders indicating that the triggering modifications were use-

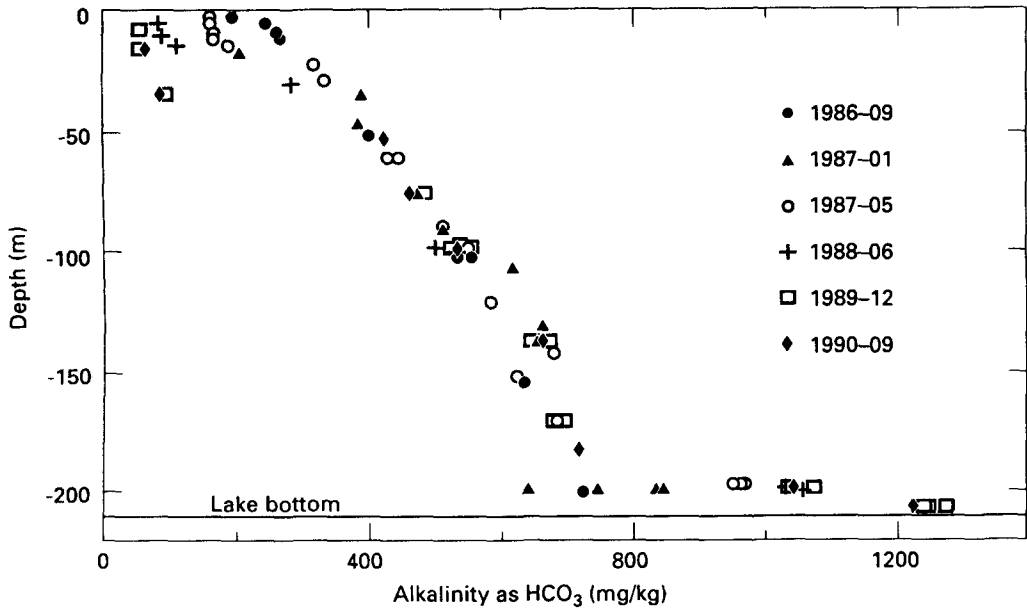FIG. 6. Alkalinity (as mg/kg HCO_3) vs depth.

ful in preventing gas loss and that the CO_2 concentrations measured in these cylinders were accurate. Comparing these concentrations to earlier results provides a measurement of the CO_2 recharge rate. Even though gas loss affected some of the earlier values, there are a sufficient number of data points that anomalously low values can be recognized and excluded from recharge calculations. For example, in cases where there are multiple cylinders from a given date and depth, the highest value should be the most reliable.

Table 2. Total dissolved gas pressure* measured by probe at various depths and stations.

Depth† (m)	Date	Station	Pressure
155	1989-12	B	0.474 ± 0.015
168	1989-12	A	0.489 ± 0.015
	1989-12	B	0.476 ± 0.015
	1990-09	C	0.475 ± 0.015
195	1989-12	A	0.709 ± 0.015
	1989-12	B	0.709 ± 0.015
	1990-09	C	0.765 ± 0.015
197	1989-12	A	0.844 ± 0.015
198	1990-09	C	0.939 ± 0.015
202	1989-12	A	0.988 ± 0.015
	1989-12	B	0.972 ± 0.015
	1989-12	C	0.995 ± 0.010
	1990-09	C	1.035 ± 0.015
206	1990-09	C	1.060 ± 0.010
211‡	1990-09	C	0.952 ± ?

* In MPa absolute.
† Depths are ±1m.
‡ Probe buried in bottom sediment.

The CO_2 concentrations obtained from all cylinders collected and analyzed to date are shown in Fig. 9. A smooth curve has been drawn through the 1990 data. At mid-depths, between 50 and 150 m, there has been little or no change with time; most samples from previous dates plot on or near the 1990 curve. The effect of recharge is apparent below 150 m depth, where values in 1990 are higher than those on all previous dates. Points marked by arrows in Fig. 9 are assumed to be invalid because of improper cylinder filling (or in the case of the January 1987 point, because of accidental opening of the entry ball valve during transport back to the laboratory). Fortunately, another sample from the same depth exists in most cases so that few data are lost. The best data to compare with the 1990 curve are the May 1987 results, because a long time interval elapsed between these two dates. Also, the May 1987 sampling was
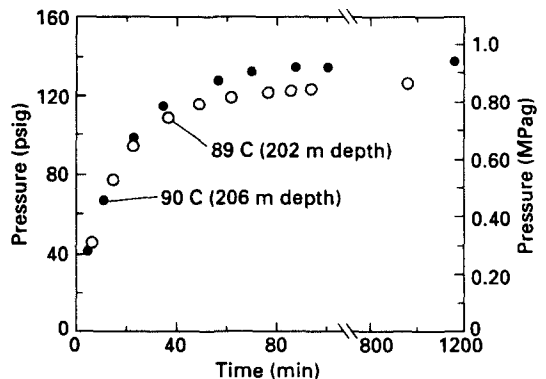

FIG. 7. Gauge pressure vs time, using gas probe at station C for two runs allowed to reach final pressures. Other pressures in Table 2 were obtained by extrapolation of similar curves after 1–2 h.

Table 3. Dissolved-gas concentrations[*] (in μmol/kg) and isotope compositions[†] (in ‰) at selected depths

Depth (m)[‡]	Date[§]	He	Ar	O_2	N_2	CH_4	CO_2	$\delta^{13}C$-DIC	$\delta^{13}C$-CH_4	δD-CH_4
76	1987-01	0.55	3.6	<1	198	111	70000	−3.8	−48.9	−241
	1989-12B	0.17	4.1	2.7	217	249	72600	−3.6	−45.0	n.a.
	1990-09C	0.21	4.7	1.1	211	299	72600	−3.8	−48.2	n.a.
99	1987-03	0.35	2.1	1.1	86.7	81.1	74900	−2.9	n.a.	n.a.
	1987-05	0.10	0.7	2.0	75.7	79.1	78800	−3.2	n.a.	n.a.
	1987-05	0.13	37.2	330	3110	162	92300	−2.9	n.a.	n.a.
	1988-06	0.29	8.1	9.9	462	249	86600	n.a.	−43.8	−235
	1989-12A	0.28	3.8	5.6	175	294	88500	−3.1	−45.6	n.a.
	1990-09C	0.29	4.8	8.3	246	309	90300	−3.4	−50.0	n.a.
137	1989-12A	0.30	3.2	3.2	158	419	128000	n.a.	n.a.	n.a.
	1989-12B	0.41	4.6	13.9	240	433	136000	−3.0	−45.6	n.a.
	1990-09C	0.42	3.6	6.1	164	430	131000	−3.2	−50.0	n.a.
171	1987-05	0.27	2.9	<0.8	166	272	132000	−3.0	n.a.	n.a.
	1989-12A	0.31	344	6670	27100	481	134000	−3.1	n.a.	n.a.
	1989-12C	0.29	2.8	6.7	177	485	138000	n.a.	n.a.	n.a.
	1990-09C	0.54	5.1	24.1	203	541	146000	−3.3	−43.1	n.a.
183	1990-09C	0.78	4.1	10.9	193	593	147400	−3.2	−46.2	n.a.
198	1987-01	1.1	4.5	2.6	146	308	159000	−3.0	−49.6	−217
	1987-03	0.64	1.4	0.7	58.0	170	146000	−3.0	n.a.	n.a.
	1987-03	1.1	2.2	1.1	159	205	177000	−3.2	−46.9	−246
	1987-05	0.40	<1.2	2.3	55.4	154	124000	−3.0	−51.2	−268
	1987-05	0.86	16.8	21.8	1560	413	193000	−3.1	n.a.	n.a.
	1987-05	0.49	<2.3	3.9	95.5	255	184000	−3.2	−44.4	−245
	1988-06	0.64	1.3	3.8	64.0	195	128000	n.a.	n.a.	n.a.
	1989-12A	1.1	2.9	2.5	101	492	203000	−3.3	n.a.	n.a.
	1989-12C	1.7	3.5	13.5	226	771	244000	−3.2	−46.9	n.a.
	1990-09C	2.0	3.8	27.1	560	876	257000	−3.5	−47.0	n.a.
206	1989-12A[‖]	1.6	1.4	1.4	150	987	281000	−3.2	n.a.	n.a.
	1989-12A[‖]	1.2	4.0	6.7	137	635	259000	−3.2	n.a.	n.a.
	1989-12B	1.3	3.5	11.3	218	1124	274000	−3.4	−46.7	n.a.
	1990-09C	1.4	5.8	17.5	457	1142	293000	−3.5	−45.8	n.a.
209	1990-09C	2.1	2.7	9.0	194	1540	299000	−3.6	−45.1	n.a.
	1990-09C	1.7	2.1	14.4	214	1730	298000	−3.4	−45.5	n.a.

[*] Analytical uncertainties are ±20% for He and ± the greater of 2% or 1 μmol/kg for the other gases.
[†] Analytical uncertainties are ±0.1‰ for $\delta^{13}C$ and ±5‰ for δD. n.a. means not analyzed.
[‡] Uncertainties are as in Table 1.
[§] Analyses from 1987-01, 1987-03, and 1987-05 are reported in KLING *et al.* (1989). Letters A, B and C refer to station locations.
[‖] Filled simultaneously.

performed from an anchored raft and included several cylinders from depths >140 m. Therefore, this profile, shown as a dashed line on Fig. 9, is well-defined.

Recharge rates and processes

Multiplying the increase in CO_2 concentration at each depth (Fig. 9) by the appropriate volume of water (Fig. 2; hypsographic curve in KLING *et al.*, 1989) gives the total increase in CO_2 as 8.6 × 10^8 moles in forty months or 2.6 × 10^8 mol/a. An analogous calculation using the May 1987 and the December 1989 data results in a very similar recharge rate, 2.5 × 10^8 mol/a. Comparing the 1990 curve to the January and March 1987 profiles yields higher recharge rates, 3.0 × 10^8 and 2.8 × 10^8 mol/a, respectively, although these profiles contain only one

depth below 150 m. An even higher recharge rate, 3.2 × 10^8 mol/a, is obtained by comparing the four points below 150 m depth in the November 1986 profile of TIETZE (1987) to the 1990 curve on Fig. 9. Thus, it appears that the CO_2 recharge rate was higher just after the gas burst than during the period between May 1987 and September 1990, despite the uncertainties in some of the CO_2 profiles. There are also uncertainties in the bathymetry. Using the hypsographic curve of NOJIRI *et al.* (1993) would reduce all the rates calculated above by nearly 15%. After accounting for this uncertainty as well as the temporal changes, the CO_2 recharge rate can still be reasonably well-constrained at 2.6 ± 0.6 × 10^8 mol/a when averaged over the first 4-a following the gas burst.

The CO_2 fluxes calculated previously by NOJIRI *et al.* (1993) and SANO *et al.* (1990) are much larger than this, by factors of 4 and 8, respectively. The discrep-
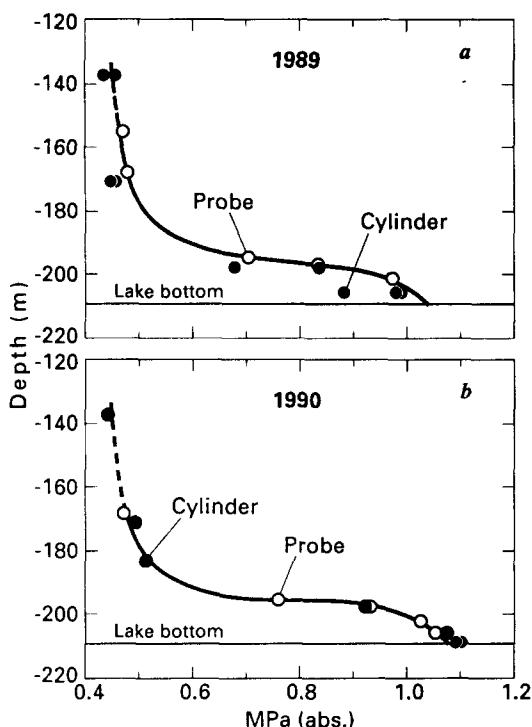
FIG. 8. *In-situ* gas pressures calculated from cylinder data compared to pressures measured using gas probe: (a) December 1989; (b) September 1990. The CO_2 partial pressures were calculated and corrected for hydrostatic effects using solubility, partial molar volume and fugacity coefficient data from WEISS (1974). *In-situ* pressures for other gases were calculated from the solubility data of WILHELM *et al.* (1977) assuming ideality. All O_2 in cylinders was assumed to be from air contamination and was used to correct N_2 and Ar values using air ratios.

ancy arises in part because these studies encompass different time intervals (November 1986–December 1988) and in part because of the assumptions made in their calculations. For example, NOJIRI *et al.* (1993) proposed a linear extrapolation of the CO_2–TDS relation from 190 m depth to lake bottom (210 m) where the water would be near saturation in CO_2. However, the CO_2 concentrations measured in 1990 just 1 m above bottom (Table 3) were less than half the calculated saturation concentration of 630,000 μmol/kg at 25°C (solubility and fugacity data of WEISS, 1974). The relation between CO_2 and dissolved solids is clearly more complex than previously thought. The data presented here show that, at 198 m depth, ionic strength has remained fairly stable since June 1988 (Table 1), while CO_2 concentrations have continued to increase (Table 3, Fig. 9). This behavior is not likely to be due to mixing of water from a deeper layer because both CO_2 and ionic strength should increase in this process. For the same reason, it is not likely to be due to diffusive transport of CO_2 from a deeper layer or double diffusive convection within the water column, the diffusion coefficients of CO_2 and most common ions being nearly equal (JOST, 1960). Further, similar behavior has occurred at 206

m depth, where CO_2 concentrations increased between the 1989 and 1990 samplings while major-ion concentrations either remained the same or even decreased slightly (Tables 1 and 3). That CO_2 and major-ion concentrations do not covary shows that no unique relation can exist between CO_2 and ionic strength (or TDS) in the bottom 20 m of the water column. Nevertheless, the proposal by NOJIRI *et al.* (1993) of CO_2 saturation at lake bottom could certainly apply to the pure recharge fluid entering the lake. This has been suggested by KUSAKABE and SANO (1991) and raises the possibility that the recharge fluid could contain free CO_2 gas, which would have important consequences for the development of water-column structure.

If the incoming recharge fluid contains free CO_2, bubbles of gas could separate from the liquid phase as it enters the lake and dissolve in the overlying water layers. Gas-phase separation could thus help explain the differing evolutions of the CO_2 and major-ion concentrations near lake bottom. However, an enormous volume flow of CO_2 gas, approximately 200 l/s (at STP; 10 l/s at lake-bottom pressure), would be required to account for the entire CO_2 recharge observed (Fig. 9). Other observations suggest that gaseous CO_2 can make up only a small fraction of the total CO_2 recharge. First, because there has been little or no increase in CO_2 concentrations above 150 m depth (Fig. 9), rising bubbles must dissolve within 60 m of the bottom. Experimental work on rates of bubble collapse (WISE and HOUGHTON, 1966) suggests that this is possible only if bubble diameters are no greater than 5–10 mm. Larger bubbles would rise to the lake surface, a process not reported at Lake Nyos. Second, horizontal variations in gas pressure and water chemistry are small even a few meters above the bottom (Tables 1 and 2). Vents of gas bubbles would be expected to disrupt the horizontal layers, and if present, must be either too weak or too widespread to generate a large, distinct plume. Vents of thermal water high in dissolved CO_2 but without abundant gas bubbles would be less likely to form a plume and better able to set up the layer structure observed in the water column. Studies of the East African Lake Kivu (TIETZE, 1978; TIETZE *et al.*, 1980), a giant analog of Lake Nyos, showed that such homogenous layers can be tens of kilometers in extent, apparently formed in response to localized inputs of thermal water. A third line of evidence is the gas composition data (Table 3). Strong streams of CO_2 bubbles would strip the less-soluble gases such as CH_4, Ar, N_2 and He from the deeper layers over time, but no obvious depletions have occurred. Concentrations of CH_4 have in fact increased.

If it is assumed that the CO_2 is entirely dissolved in the recharge fluid, the CO_2 recharge rate can be used to calculate a range of water flow rates and temperatures. The minimum CO_2 concentration in the incoming fluid is apparently 0.30 mol/kg, the highest measured value at 209 m depth (Table 3). A CO_2 flux
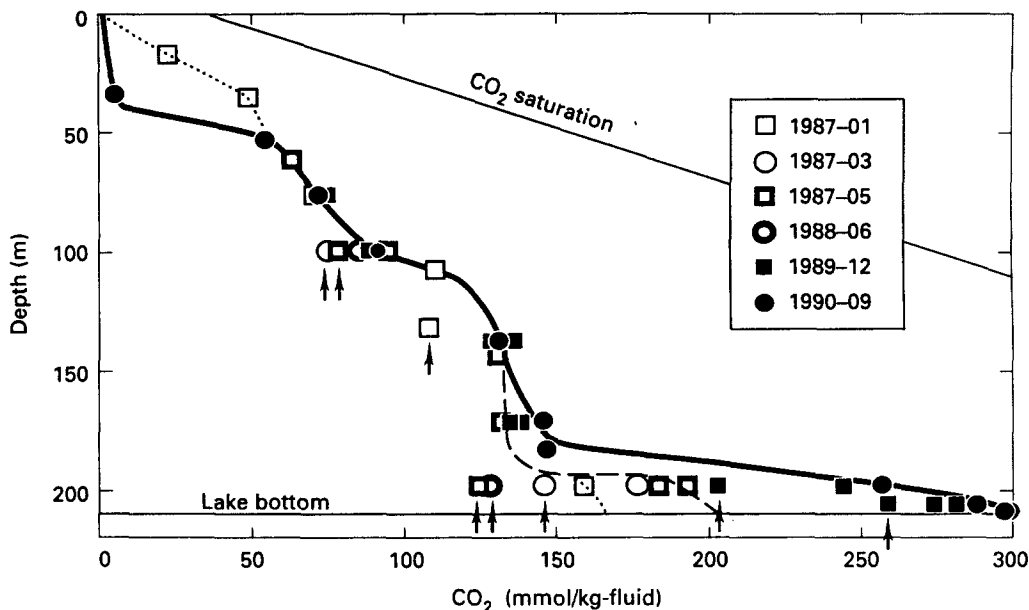
FIG. 9. Concentration of CO_2 in cylinder samples from all dates and depths. The recharge rate is calculated by comparing the September 1990 curve (solid line) to the May 1987 curve (dashed line) or the January 1987 curve (dotted line) at depths below 150 m. Points marked by arrows are presumed to be low due to cylinder failure (see text) and are not used in the recharge rate calculations. Surface loss of CO_2 is shown by the profile evolution between 0 and 50 m depth. Field measurements of pH and alkalinity have helped define the profiles in this region of lower pCO_2 values. The "CO_2-saturation" line represents the maximum possible concentration of CO_2 in solution as a function of total pressure, 89 kPa atmospheric $+9.76$ kPa/m depth.

of 2.6×10^8 mol/a would then require an input of 8.7×10^5 m³/a (27 l/s) of water. If the increase in heat content below 100 m depth (8.1×10^9 kcal between May 1987 and September 1990) were due entirely to transport by the recharge water, the water temperature would be 26.1°C, slightly warmer than the 24.9°C actually measured at 209 m depth (Fig. 5). Repeating the calculations for a recharge fluid that is CO_2 saturated at the lake-bottom pressure of 2.1 MPa (21 atm) yields a temperature of 28.6°C, a CO_2 concentration of 0.57 mol/kg and a flow rate of 4.7×10^5 m³/a (15 l/s). These ranges of temperature and flow rates are close to the 30.4°C and 13 l/s values calculated by Nojiri *et al.* (1993) based on the assumption that the recharge fluid is saturated with amorphous silica. The agreement is very good given the difference in the two approaches. Recharge is thus a low-temperature process, more consistent with the existence of a low-temperature source of CO_2 than a near-surface magma chamber or high-temperature hydrothermal system.

The range of temperatures and flow rates could be larger if a significant fraction of the CO_2 recharge consists of free gas. The heat of solution of CO_2 gas in water is considerable (-4.7 kcal/mol) while the heat capacity is only 9.1 cal/mol°C. When CO_2 dissolves in water, there is a rise in water temperature that is mainly due to the heat of solution, even if the CO_2 is initially much hotter than the water or performs expansion work. If the recharge fluid contains free CO_2, part of the hypolimnetic heat increase would

derive from the heat of solution and part from fluid transport. In the extreme case, where it is assumed that all the CO_2 recharge consists of free gas, the increase in hypolimnetic heat content due to the heat of solution would be 4.0×10^9 kcal, about half the increase actually measured. Required recharge fluid temperatures would then be 1–2°C cooler assuming the same range of flow rates as calculated above.

There may be important temporal variations in recharge temperatures and flow rates as well. The ranges reported here are average values for the period between May 1987 and September 1990. They mainly serve to show that recharge temperatures are low and do not imply that the recharge fluid has a fixed composition. In fact, a fixed recharge-fluid composition is difficult to reconcile with the behavior of the major-ion chemistry in bottom waters. As noted previously, increases in CO_2 concentration (Table 3) are no longer accompanied by increases in ionic strength (Table 1). The data from 198 m depth, displayed in Fig. 10, suggest that between the May 1987 and June 1988 samplings, the rapid influx of ionic species into bottom waters ceased. The cessation may not have been as abrupt as implied in Fig. 10, but it appears to have occurred during the same time interval as the drop in the rate of CO_2 recharge and to require a change in recharge-fluid composition rather than simply a change in flow rate.

In a simple 2-component mixing model where the recharge fluid continually vents into the more-dilute hypolimnetic water, concentrations of CO_2 and ionic
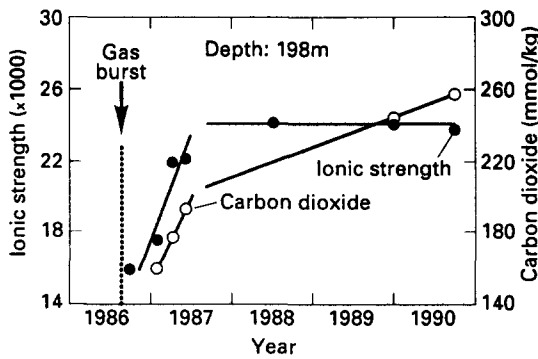
FIG. 10. The evolution of CO_2 and ionic strength with time at 198 m depth. Concentrations of CO_2 are taken from Table 3 assuming that the highest value on each date is the most reliable (see text). The cylinder from June 1988 is assumed to have lost CO_2 based on Fig. 9.

species in the hypolimnion would be expected to increase after the gas burst but then level off as some steady-state conditions were reached between inputs and transport through the water column. However, the approach to steady-state should be gradual, rather than sudden as has been the case with ionic strength (Fig. 10). Also, the concentrations of all chemical species including CO_2 should completely level off at the same time, even if recharge fluid flow rate is variable. After June 1988, CO_2 recharge continued at a reduced but easily measurable rate, while modifications in water chemistry have become restricted to changes in the relative proportions of the various ions and possibly, slight decreases in ionic strength (Table 1). Mineral saturation could play an important role in controlling certain ion concentrations; for example, authigenic siderite has been reported in the lake sediments (BERNARD and SYMONDS, 1989). However, the abrupt flattening in the ionic strength trend (Fig. 10) was certainly not caused entirely by mineral saturation. Bottom waters remain undersaturated with respect to calcite (TUTTLE et al., 1992), and yet Ca as well as Na and Mg concentrations level off or drop slightly after June 1988. Given the topographic relief around Lake Nyos (Fig. 2), the mixing model should contain a third component; that is, a dilute meteoric fluid flowing into the lake bottom from the granitic crater walls as suggested by TUTTLE et al. (1992). A sudden increase in the flow rate of this dilute fluid, or decrease in the flow rate of the recharge fluid, could cause rate changes similar to those in Fig. 10. However, between May 1987 and June 1988, there were no reports of large earthquakes or sudden changes in lake level that could significantly affect groundwater flow in the area. In addition, the relative flow rates of the recharge fluid and the meteoric fluid after June 1988 would have to balance in such a way as to cause only small changes in major-ion concentrations in bottom waters.

It is difficult to account for a change in the composition of the recharge fluid, which is similar in many respects to soda-spring type fluids. Soda springs result when groundwater contacts a potent source of CO_2 and the acidic solution chemically attacks the surrounding rock to become mineralized. Many such springs are known to exist in Cameroon (MARECHAL, 1976), and in general, soda springs are stable features than can persist for millennia (see e.g. BARNES et al., 1984). The most-probable scenario is that the recharge fluid is mainly hypolimnetic water involved in a sub-bottom convection system: sinking down through bottom sediments; acquiring CO_2, heat, and ionic species within the underlying diatreme; then rising up through faults or fractures to vent back into the lake. During the gas burst when dilution of the hypolimnion occurred, this sub-bottom convection system, labeled path A in Fig. 11, would still contain pre-burst hypolimnetic water. As long as this more-concentrated water continued to vent, it caused the rapid increases in bottom-water ionic strength and CO_2 concentration observed through May 1987 (Fig. 10). By June 1988, the water in the convection system had been replaced by more-dilute, post-burst hypolimnetic water (that is, post-burst water had completed one cycle along path A in Fig. 11), and increases in ionic strength tailed off and became undetectable. Circulating water continued to transport CO_2 into the lake, but at a reduced rate, similar to the 2.6×10^8 mol/a calculated above from the May 1987 and September 1990 data. This presumably represents the actual supply rate of CO_2 within the diatreme.

The diatreme is thought to contain rocks similar to those found in the ejecta, including alkali basalt and ultramafic xenoliths (described by SCHENKER and DIETRICH, 1986; LOCKWOOD and RUBIN, 1989). Weathering of the mafic minerals in these rocks by the recharge fluid is responsible for the unusual major-ion chemistry in bottom waters (TUTTLE et al., 1992) and some CO_2 is no doubt extracted from the rocks during the weathering process. However, the release of 2.6×10^8 mol/a of CO_2 entirely by weathering should be accompanied by a major release of ionic species into the recharge fluid. The stable major-ion concentrations after June 1988 (Fig. 10; Table 1) show that there is little net transport of ionic species by the recharge fluid as it cycles through the closed loop and that weathering is a slow process compared to the rate of CO_2 transport. Thus, the recharge fluid must encounter a pocket of free CO_2 during the closed-loop cycle. This CO_2 reservoir must extend to very shallow levels within the diatreme given that the closed-loop water-circulation time is only ~1 a, as shown by the time required to flush the more-concentrated pre-burst water out of the loop (Fig. 10). The reservoir could contain the CO_2 as liquid, a high-pressure gas, or a supercritical fluid depending on P-T conditions (KUSAKABE and SANO, 1991; NOJIRI et al., 1993). Such reservoirs are known and can be large. A reservoir of fluid CO_2 near a Recent Australian maar has been drilled to produce CO_2
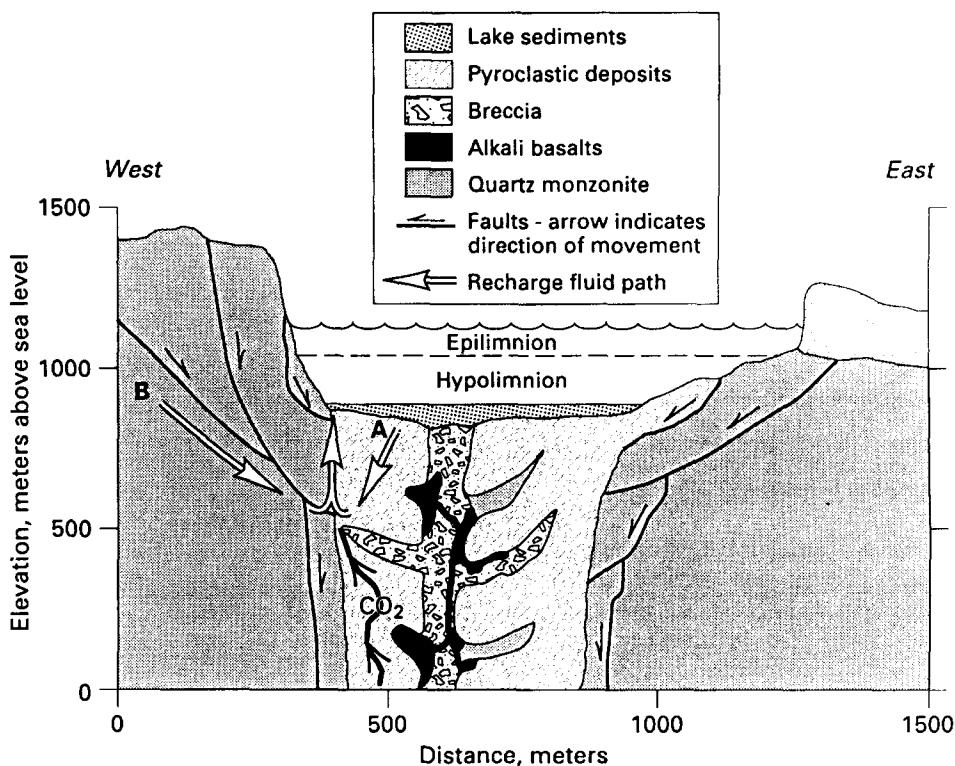
FIG. 11. Cross-section of Lake Nyos and the underlying diatreme (modified from TUTTLE *et al.* 1987). Hypothetical recharge fluid flow paths are shown as "A" for closed-loop and "B" for external source. Reservoir containing free CO_2 extends up to the zone of contact with groundwater.

commercially at a rate of 6.3 × 10^{8} mol/a (CHIVAS *et al.*, 1987), more than double the recharge rate at Lake Nyos.

A similar cycling of lake water has been proposed for the crater lake in the active Poas volcano, Costa Rica, where sub-bottom convection is driven by magmatic heat (ROWE *et al.*, 1990). At Lake Nyos, heat is probably only part of the driving force behind convection given that in bottom water, the slowly rising temperatures (Fig. 5) do not offset the density increases caused by the rising concentrations of dissolved CO_2 (Fig. 10). Additional force could be provided by buoyancy. If the CO_2 rapidly enters the water at the bottom of path A (Fig. 11), the density of the water would be greatly reduced and remain so until the CO_2 completely dissolved. During this time, buoyant ascent of the fluid could provide considerable driving force for convective motion. As noted earlier, complete dissolution of the CO_2 is likely to occur by the time the recharge fluid vents into the lake, and any residual gas bubbles must dissolve within the lower 60 m of the water column.

Additional samples and more-complete chemistry profiles would help test this closed-loop hypothesis as well as the constancy of the CO_2-supply rate. Some additional evidence of a closed-loop system is given by the bottom-water Ar values (Table 3). The concentrations of this inert atmospheric gas in the local meteoric water or in groundwater directly derived

from local meteoric water should be 11–13 μmol/kg (solubility data of WILHELM *et al.*, 1977), and values in this range are found in the hypolimnia of other stratified crater lakes in Cameroon (KLING *et al.*, 1991). The Ar concentrations in Lake Nyos are much lower because of losses during the 1986 gas burst but would be expected to increase with time in the deepest layers if the recharge fluid were primarily from a source external to the lake (e.g. path B in Fig. 11).

Gas burst recurrence

As hypolimnetic CO_2 concentrations rise, the danger of another gas burst increases, but it is difficult to establish a precise relation between recharge rate and the probability of a gas burst. If CO_2 recharge were to remain constant at 2.6 × 10^{8} mol/a, it would require ~160 a to saturate Lake Nyos given the present-day CO_2 concentrations (Fig. 9) and assuming the development of a unique mixing and stratification pattern allowing CO_2 concentrations to just equal saturation at every depth. As the stratification pattern of the lake continues to evolve, the depth of seasonal mixing will determine the shallowest level at which gas build-up can occur. To date, some mixing has been observed down to 50 m depth (Fig. 6). If this is where a stable chemocline develops,

the time required to saturate the lake below this level should be ~140 a, again assuming a constant recharge rate. In a simplistic model of the gas-burst process where dissolved CO_2 slowly builds up to saturation pressure and is suddenly released, this 140-a period would represent the recurrence interval. However, there is no proof as yet that complete saturation of the hypolimnion can ever occur, and certainly it is not required before a gas burst. Regardless of recharge rates, another gas burst could occur any time there is sufficient mechanical input, such as a landslide, to move a large parcel of bottom water into shallower levels where dissolved gas pressure exceeds total pressure. Bubble formation would then greatly reduce fluid density, and buoyancy forces would render additional mechanical input unnecessary. The condition required for a gas burst to ensue is that depressurization of adjacent water be sufficient to sustain circulation, causing degassing of a significant fraction of the hypolimnion (see discussion by KANARI, 1989). Currently (as of 1990), water at 209 m depth, with a dissolved gas pressure of 1.1 MPa (11 atm) (Fig. 8b), would have to rise to depths of <110 m for bubble formation to begin. Assuming a closed-loop recharge system, CO_2 concentrations in the recharge fluid will probably increase as a function of time (i.e. CO_2 is incrementally added to the fluid as it circulates through the diatreme). In this case, the very deepest layers in the water column are likely to saturate well before the mid-depths, with there being no net influx of water into the lake bottom to force overlying water layers upward. Before additional constraints can be placed on a theoretical recurrence interval, factors other than CO_2 recharge rate must be considered, a few of which are discussed here.

Vertical mixing is extremely limited given the large density gradient in Lake Nyos, but increasing temperatures in bottom water (Fig. 5) work to lessen the density gradient. A fluid with bottom-water composition (e.g. at 206 m depth, Tables 1 and 3) would need to reach 36°C to become less dense than 22°C surface water. This is above the calculated range of recharge fluid temperatures and well above present bottom-water temperatures, which appear to be leveling off near 25°C. Thus it is unlikely that temperature alone will upset the density stratification. However, increased temperatures have a large effect on dissolved gases, reducing the solubility of CO_2 by ~2.5%/°C. By weakening the density stratification and by decreasing the CO_2 storage capacity, higher bottom-water temperatures increase the danger of a gas burst recurrence.

There are surface losses of CO_2, as shown in Fig. 9. Between January 1987 and September 1990, an estimated 1.3×10^9 moles of CO_2 (3.5×10^8 mol/a) were gradually lost from the surface layers. This is greater than the amount gained by bottom layers and so despite continuous recharge, Lake Nyos contains less gas at present than soon after the 1986 burst. This high surface flux is not likely to represent a long-term

rate but is a consequence of the unusual lake structure following the 1986 partial mixing (pCO_2 in surface waters much higher than in the atmosphere). As the chemocline becomes more pronounced, the loss of CO_2 from the lake surface should decrease due to the small transport rate across this interface. However, unless the transport rate across the chemocline becomes trivial compared to the recharge rate, the effect of surface losses will be to slow the rate of gas build-up in the hypolimnion and increase the recurrence interval.

In terms of concentration, gases other than CO_2 make up <1% of the total dissolved gas (Table 3; air-corrected), but because of lower solubilities, N_2 and CH_4 add noticeably to the total gas pressure. At 209 m depth they currently contribute 0.14 MPa or 13% of the total gas pressure of 1.1 MPa (calculated from Table 3, Fig. 8). If water from 209 m depth were depressurized in a Rayleigh process, the first small bubbles would be ~87% CO_2, 13% N_2 and CH_4, in proportion to their respective pressures. The gas phase would get progressively richer in CO_2 until, at complete depressurization, it would be ~99% CO_2, 1% N_2 and CH_4, in proportion to initial concentrations in solution. Although rapid degassing might only approximate a Rayleigh process, N_2 and CH_4 could be instrumental in initiating a gas burst but make up a trivial proportion of the gas released. The effect of these gases will be to lower the amount of CO_2 that the hypolimnion can store and to increase the danger of recurrence.

It is alarming that bottom-water CH_4 concentrations are increasing at a faster rate than CO_2 concentrations (Table 3). Much of the CH_4 increase results from biogenic production. This is evident because CH_4 concentrations have also increased at mid-depths, well above the bottom 40–60 m of the water column where recharge effects on water chemistry, temperature, and CO_2 are seen (Figs 5, 9; Table 1). Also, one ^{14}C analysis has been completed. The CH_4 in sample 12-89B from 206 m depth (Table 3) contained 41% modern C, probably resulting from bacterial fermentation of sinking organic matter. The percentage of the CH_4 that is biologically produced is certainly higher than this. Any CH_4 produced by bacterial CO_2 reduction, for example, would contain only the 1–2% modern C present in the DIC (TUTTLE et al., 1987). Note also that the $\delta^{13}C$ values of the CH_4 show no trends with depth or time (Table 3). Either the biological processes produce CH_4 that is isotopically similar to that brought in by the recharge fluid or the recharge fluid does not introduce significant amounts of CH_4. The $\delta^{13}C$ values are heavy for biogenic CH_4 (cf. WHITICAR et al., 1986), but similar values have been reported for Lake Kivu (DEUSER et al., 1973) and Lake Monoun (EVANS et al., 1989). If CH_4 concentrations continue to increase at present rates, they could soon have a large effect on total gas pressure in the hypolimnion. It is possible, however, that biogenic CH_4 production has been enhanced by

nutrient redistribution during the gas burst as has been suggested for Lake Monoun (EVANS *et al.*, 1989). If this is the case, methanogenesis may decrease with time.

CONCLUSIONS

Using anchored sampling stations and a new gas probe, it has been shown that horizontal variations in CO_2 concentrations, like those in temperature and water chemistry, are small in the bottom waters of Lake Nyos. The rate of horizontal mixing must be very rapid compared to the rate of vertical mixing, and profiles at a single site can therefore be used to calculate fluxes for the entire central basin. The CO_2 recharge rate of 2.6×10^8 mol/a is much smaller than previous estimates (SANO *et al.*, 1990; NOJIRI *et al.*, 1993) but may better represent the long-term rate of supply from the underlying diatreme. It is certainly large enough to support the hypothesis that long-term build up of CO_2 in the stratified hypolimnion can occur, a critical part of the limnological theory of the gas burst. The resulting 140-a period required to achieve saturation in the hypolimnion favors the possibility that previous gas bursts have occurred at Lake Nyos even though the age of the crater may be only 400–500 a (LOCKWOOD and RUBIN, 1989). However, it is only a crude estimate of the recurrence interval.

The CO_2 recharge rate is of immediate value in plans to artificially pipe the gas out of bottom waters (HALBWACHS *et al.*, 1990). The piping rate should at least equal the recharge rate to be effective in preventing gas accumulation. However, the rapid horizontal mixing should allow the entire bottom area to be degassed by pipes installed in one location. Proposed remedial actions also include lowering lake surface level 40 m to the base of the natural spillway, which is structurally weak and could fail causing massive flooding (LOCKWOOD *et al.*, 1988; SCHUSTER and LOCKWOOD, 1991). Hydrostatic pressures would be reduced by 0.4 MPa in this process; meaning that, should the dam fail prior to the proposed gas removal process, dissolved CO_2 pressures would be closer to saturation values. This can be visualized by lowering the "CO_2 saturation" line 40 m on Fig. 9. At no point on the 1990 profile would CO_2 concentrations be saturated, but a small disturbance of the water column could more easily initiate a gas burst. Also, a gas eruption could begin beneath the lake floor if pressure is reduced by either dam failure or intentional lake drainage (see TIETZE, 1987). This is especially a threat if a CO_2 reservoir exists only a short distance below lake bottom as suggested by the closed-loop recharge model, and the danger would remain even after completion of gas removal from the hypolimnion. On the other hand, if the CO_2 reservoir does extend to shallow depths, it may be possible to drill into the reservoir and release the

trapped gas directly to the atmosphere, by bypassing the existing recharge system. If successful, this procedure (discussed by SCHENKER and DIETRICH 1986) could eliminate the problem of gas buildup in the hypolimnion as well as the danger in lowering the lake level. Exploratory drilling should be attempted as part of the hazards reduction plan.

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