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TOWARDS A JOINT GLOBAL OCEAN FLUX STUDY: RATIONALE, OBJECTIVES, PLANNING, IMPLEMENTATION

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ABSTRACT

The oceanic inventories and fluxes of Carbon are of major importance for the global C cycle in past, present and future. For the very near future the oceans act as moderator for the atmospheric CO_2 increase due combustion of fossil fuels. The current understanding of ocean/atmosphere exchange and fluxes into the ocean interior is however modest and considerable more insight is required in the way the ocean controls the C cycle. The global scale of the problem and the intricate interactions between physical, biological, and chemical forces warrant both an international and a truly multidisciplinary approach. The important role of marine biota as well as its interactions with many other (trace) elements deserve major attention. For proper global quantification of the C cycle large scale observation of the ocean surface from satellite is the only available tool. Objective of the ICSU/SCOR/JGOFS program is

'To determine and understand on a global scale the time varying fluxes of carbon and associated biogenic elements in the ocean, and to evaluate the related exchanges with the atmosphere, the sea floor and continental boundaries'.

Planning for 1989-1999 is focusing on (1) Process studies in selected areas; (2) Time Series observations to be continued or newly started; (3) Global Survey largely through satellite observation. The merits of selected study areas in regions as the Antarctic Ocean, the Pacific Basin, the Northwest Indian Ocean and the North Atlantic Ocean are being exploited. Implementation is now well underway with the North Atlantic Pilot Study where ships from six countries participate in a study of the temporal and spatial shifts of C fluxes related to the spring and autumn plankton blooms.

INTRODUCTION

The Global Ocean. - The ocean covers most of the surface of the globe. Its metabolism, i.e. the internal cycling and boundary exchanges of chemical constituents of the ocean, has a major impact on the atmosphere and the terrestrial biosphere. One example is the cycle of Carbon, the prime building block of all biomass (Figure 1). The seawater reservoir contains about 40,000 x 10^{15} grams or 40,000 Gigaton Carbon (largely as the bicarbonate ion: $HCO3^-$) and its annual exchange rate with the atmosphere is on the order of 100 GtC per year. The atmospheric reservoir itself contains only about 700 GtC as carbondioxide (CO2) and exchanges with a rate of about 70 GtC/yr with the terrestrial biosphere. The latter contains about 2000 GtC, of which some 70 percent, say 1400 GtC in the soil (humus) with the remaining 600 GtC aboveground as plants.

Throughout geological time there have been variations in the *GEOSPHERE / BIOSPHERE* system, including shifts in the above inventories and fluxes of the C-system. For example, records over the past 160,000 years derived from Antarctic ice cores (Figure 2) show atmospheric CO2 variations between about 120 x 10^{-6} atm and 270 x 10^{-6} atm, the latter correlating with variations in surface temperature (BARNOLA ET AL., 1987). The apparent cause/effect relationship of this glacial / interglacial CO_2 shift (NEFTEL ET AL., 1988) is the subject of considerable debate (KNOX & McELROY, 1984; BROECKER, 1982, 1987; SARMIENTO & TOGGWEILER, 1984; MIX & FAIRBANKS, 1985; BOYLE, 1986, 1988). Yet there is no doubt that the oceans, through variations in circulation, biological productivity, nutrient regime, etc., play a crucial role (SUNDQUIST & BROECKER, 1985; MOORE & BOLIN, 1986; BOLIN ET AL, 1986; BROECKER ET AL, 1987; SARNTHEIN ET AL., 1988; BERGER, SMETACEK & WEFER, in press).

CO2 and global warming. - During the past century mankind has begun to influence the geochemical system on a global scale. With respect to the C cycle there is the well documented emission of CO2, exponentially increasing in time until about 1973, to an annual rate which is currently more or less stable at about 5-6 GtC/yr (Figure 1). Time series observations (KEELING ET AL., 1976a,b) have shown this to lead to an increase in atmospheric CO2 levels from about 270 x 10^{-6} atm in the nineteenth century through about 315 x 10^{-6}atm in 1958 to about 350 x 10^{-6} atm in 1988 (Figure 3). This level is still increasing, and already

much higher than ever before in the past 160,000 years (Figure 2). Mankind has truly ventured into its 'greatest' experiment ever (REVELLE as cited by BRYAN, 1986). The growing envelope of atmospheric CO2 may, through enhanced heat absorption, lead to a global warming trend, the so-called 'greenhouse effect', possibly also accompanied by melting of ice caps and a general sea level rise (TYNDALL, 1863; CALLENDAR, 1938; NATIONAL ACADEMY OF SCIENCES, 1983; SCHLESINGER, GATES & HAN, 1985; TRABALKA & REICHLE, 1986; FERGUSON, 1988). The likelyhood of such CO_2 induced global warming is significant, also from judging the aforementioned ice-core records, but accurate prediction of the timing and magnitude of such event is not yet possible. Keen observers have indeed reported an increase in global surface air temperature of about 0.5 °C over the past century (JONES ET AL., 1988). Yet the underlying database as well as the statistical (long vs. short term, signal vs. noise) significance of such trend appears to be rather debatable (e.g. HARE, 1988). More importantly the suggested trend may also be ascribed to the fact that the deep ocean (with a response time of centuries) and atmosphere are probably still adjusting (BRYAN, 1986) from the 'Little Ice Age' (GROVE, 1988). Latter event was most severe in the early 17th century (Figure 4), a period also known for the unusual high number of 'ice-scapes' with skating figures created by Flemish and Dutch master painters (Figure 5).

Comparison of emission rates derived from fossil fuel (coal, oil, gas) statistics (NATIONAL ACADEMY OF SCIENCES, 1983) with the observed CO2 increase has learned that only about half (estimates vary between 25-60 percent) of the fossil fuel derived CO2 is retained in the atmosphere. The other half (at its extremes 40-75 percent) is presumably taken up by the oceans. The uncertainty derives from the fact that net CO2 emissions due to changes in terrestrial biomass (deforestation, agriculture) are very difficult to assess. Estimates vary between extremes of -1 and +4 GtC/yr, corresponding to a small increase or larger decrease of terrestrial biomass respectively. Nevertheless it is obvious that the oceans serve as major sink for atmospheric CO2. Through dissolution of CO2 in the upper mixed layer (about 75m depth) of the ocean as well as biological C-fixation (photosynthesis) the surface ocean can take up about 2-3 GtC/yr, with an upper extreme estimate of 7 GtC/yr (Figure 1). Biological activity may well be the most crucial factor controlling the actual rate of CO_2 uptake. This net amount is only a fraction of the gross 100 GtC/yr annual atmosphere / upper ocean exchange. The physico-chemical uptake of CO2 in surface waters is

further predicted to actually decrease with increasing CO2 levels (Revelle factor, TAKAHASHI ET AL., 1980). The actual effectiveness of the net ocean sink is otherwise not well known anyway and also varies from year to year. For example, significant changes in atmospheric CO2, were noted during the El Nino - Southern Oscillation events in the South Pacific Ocean (BACASTOW, 1976, 1977; KEELING & REVELLE, 1985). The exchange of CO2 between atmosphere and ocean is clearly not in steady state but subject to dynamic variability throughout the frequency domain, further enhanced by the recent changes generated by industrial CO2 emissions and dramatic clearing of tropical forests.

Nevertheless the excess, say 2-3 GtC/yr, CO2 taken up in the surface ocean would eventually be carried downward by both water circulation (sinking rate of CO2-saturated water at the poles) as well as biogeochemical processes. We are currently however not able to quantify whether the physical and chemical processes alone (ocean heating, cooling, mixing, CO2 dissolution) or the biological processes (photosynthesis and respiration) predominate the contemporary oceanic sink, both with respect to capacity as well as actual uptake rate of CO_2 (BREWER, 1986). Clearly a better understanding and quantification of latter processes is crucial for modeling the atmospheric CO2 content and ensuing 'greenhouse' effects. Scenarios for the latter (HANSEN ET AL., 1984; WASHINGTON & MEEHL, 1984; WETHERALD & MANABE, 1986; SCHLESINGER, 1986; WILSON & MITCHELL, 1987; SCHLESINGER & MITCHELL, 1987; SCHLESINGER & ZHAO, 1988) are in great need of proper oceanographic constraints (SCHLESINGER, 1988) in order for them to serve as reliable tools for policy decisions on matters as deforestation, curtailing CO2 emissions and anticipation of possible sea level rise (FERGUSON ET AL., 1988).

Other 'Greenhouse' Gases. - The biosphere and its perturbations are however more complicated than the C system alone. For example mankind also introduces other 'greenhouse' gases into the atmosphere, notably ammonia (NH3), nitrous oxide (N2O), ozone (O3), methane (CH4) and ChloroFluoroCarbons (CFC's or freons). Their combined effect on the global temperature is not very well understood, but may well be comparable to the CO2 effect alone. For methane, another component in the C cycle, more recent work has shown an appreciable annual increase of atmospheric CH_4 burden (FRASER ET AL, 1981; RASMUSSEN & KHALIL, 1981[a,b], 1984) in both hemispheres. This is consistent also with an increasing trend dating as far back as about 1580 A.D. as observed in ice cores (CRAIG & CHOU, 1982). In some respects the methane story is

not dissimilar to that of CO_2 and in the various greenhouse scenario's possibly 30 % of the predicted warming rate may be ascribed to the methane increase alone (SENUM & GAFFNEY, 1985).

All these gases, except the CFC's, have always existed in natural cycles and complicated feedback mechanisms are conceivable or known to exist. For example, anthropogenic atmospheric emissions (NH_3; NOx) as well as agricultural inputs into rivers (fertilizers) may through nutrient elements nitrogen (N) and phosphorus (P) yield higher rates for biological C-fixation in the upper ocean, i.e. enhanced removal of CO2 from the atmosphere. For ChloroFluoroCarbons (freons) (RASMUSSEN and KHALIL, 1986; WIGLEY, 1987) we know that they are also taken up by the oceans (and in fact employed as water mass tracers by oceanographers, BULLISTER & WEISS, 1983; WEISS ET AL., 1985; WALLACE & LAZIER, 1988), i.e. the ocean may again act as a moderator with respect to both global warming as well as the freons / ozone interaction.

Strictly natural components cannot be overlooked either (ANDREAE, 1986). Biological activity in the upper ocean yields DiMethylSulfide (DMS) which is partially emitted into the atmosphere. Atmospheric oxidation leads to sulfate aerosols which serve as nuclei for cloud formation, thus influencing climate through both albedo (reflection of incoming sunlight) and heat insulation. Other natural gaseous compounds like CarboonylSulfide and various methylhalides are also released from the ocean into the atmosphere, where they affect heat budget and ozone layer.

Notwithstanding their importance in the global JGOFS study we will in this paper further refrain from discussion of all these 'other' greenhouse gases, limiting ourselves instead to a 'case-study' of just CO_2. For this purpose we will first focus on the relation of the oceanic C cycle to CO_2 and secondly highlight some current issues of relevance for the marine C system.

CARBON POOLS AND PATHWAYS IN THE OCEANS

Throughout most of the world ocean there exist both a seasonal and permanent thermocline at depth intervals typically near 50-150m and 200-800m respectively, which are characterized by strong temperature gradients. The corresponding density gradients largely prevent vertical mixing of water and its dissolved constituents between the overlying surface ocean and the below deep ocean interior.

Only the surface mixed layer, with a typical depth in the 25-250m range driven by wind mixing, is capable of exchanging gases like CO_2 with the atmosphere. The direction of CO_2 gas exchange is driven by in themselves relatively straightforward equilibrium considerations (under- or oversaturation); the actual rate of exchange by interaction of complex processes and concepts like wind induced mixing, bubble injection, surface film thickness (O'BRIEN ET AL., 1986).

The very same surface layer is also known as the euphotic zone, the depth interval (down to about 1 % light penetration) where photosynthetic C-fixation (primary production) by algae is responsible for the massive conversion of dissolved inorganic bicarbonate into plant organic matter. This leads to lower values for DIC in the surface ocean (Figure 6), hence allowing further physicochemical dissolution of atmospheric CO_2. Virtually all (say 90-99 %) of this carbon Gross Primary Production (GPP) is converted through grazing by zooplankton, microbial degradation and other processes within the complex food web of the upper ocean, yielding both Dissolved Organic Carbon (DOC) as well as complete recycling (mineralization, respiration) of organic C back again into the Dissolved Inorganic Carbon pool (Figure 7).

However some of the plant / animal material or Organic Carbon (POC as well as DOC), will escape mineralization and is exported to deeper waters. This 'New Production' term (DUGDALE & GOERING, 1967; EPPLEY & PETERSON, 1979; EPPLEY, 1988) is roughly one order of magnitude less than the Gross Primary Productivity but represents the biologically mediated net flow of C from the atmosphere into the deep ocean. The larger size fractions in the particle spectrum settle down out of the euphotic zone into the deep ocean. This vertical flux of organic particles is one major pathway for transfer of C from the atmosphere / surface ocean into the interior, the 'biological pump' for removal of atmospheric CO_2. Concurrently biologically essential elements (N, P, Si) are also being removed to such extent that very little remains available for plant growth. At 'near-zero' levels of nitrate, ammonia, phosphate or silicate the surface ocean productivity is typically limited by N availability. Only through replenishment of N by physical mixing from deeper waters (or possibly from atmospheric input) can the new production flux be sustained.

Another pathway is the sinking of surface water, by means of intense winter cooling. Latter process occurs seasonally, only in the northern North Atlantic (Norwegian and Greenland Seas, Labrador Sea) and the Weddell Sea off Antarctica. Through its tendency to equilibration with

the atmosphere the very cold water acts as a sink for inorganic CO_2, moreover it also carries down an inventory of DOC.

Sluggish circulation within the dark abyss slowly fills in the about 4000 m deep basins of the major oceans; the mean 'age' of the deep water is about 500-1500 years, depending on whose ^{14}C based model calculations one believes (STUIVER ETAL., 1982; BROECKER & PENG, 1982). Throughout such time scale most of the DOC will be converted to DIC, also the settling large particles will be remineralized to DIC (while settling but also at the seafloor). From the vertical and interoceanic distribution of various dissolved tracers (nitrate, phosphate, $\Sigma CO2$, $\delta^{13}C$, etc.) in the world ocean we know that the net flux of organic matter originating from the surface ocean is remineralized within the deep sea and at the seafloor. We are not certain though whether this organic matter is brought into the deep sea largely as particles settling from the euphotic zone, or mostly as DOC carried along with the general water circulation.

The water balance of the surface ocean is restored by a mean upward advective flow or upwelling in the order of 4 m per year. In certain locales, e.g. the equatorial Atlantic, more intense upwelling is noted from excess CO_2 contents of the water as compared with the atmosphere (Broecker & Peng, 1982, see also Figure 5). Some of this excess is fixed by photosynthesis, some escapes into the atmosphere, as was concluded from $^{13}C/^{12}C$ analyses of atmospheric CO_2 (KEELING ET AL., 1984). The short mixing time of the atmosphere makes each hemisphere fairly homogeneous with respect to CO_2; the ocean 'breathes' CO_2 by inhaling at the poles and exhaling at the equator. Above we found this ocean/atmosphere CO_2 exchange to be massive (about 100 GtC/a) compared with the modern fossil fuel CO_2 additions (about 4 GtC/a) (Figure 1).

BIOLOGICAL PRODUCTION

Within the context of a global study of the marine C cycle the marine ecosystem deserves and obtains a great deal of attention (e.g. U.S.GOFS Report 7, 1988; NERC, 1987; NETHERLANDS SSG, 1988; MILLS ET AL., 1987). Within the modest context of this paper we forced ourselves to largely ignore the important intricacies of the marine foodweb; limiting ourselves instead to a simplified black box approach.

The biological fixation of C within the euphotic zone is regulated by physical forcing: light regime, temperature as well as the physical

mixing (e.g. tides, wind, internal waves) required to replenish the limited supply of essential (typically N) nutrients (DICKEY & SIEGEL, 1988). In different areas of the worlds ocean (e.g. euphotic zone of North Atlantic, North Pacific, Antarctic Circumpolar waters) very different ecosystems with their own dynamics (diel to seasonal to interannual variability) do exist. The ensuing export or New Production term varies accordingly, as does the 'efficiency' or

f-ratio = New Production / Gross Primary Production.

In other words, the fluctuations or outright episodic nature of the physical forcing and biological response yield considerable patchiness of the overall marine ecosystem, largely preventing adequate in situ sampling on a global scale. However biochemical tracers (e.g. nutrients, dissolved O_2) in the deep ocean provide a 'time-and-space' averaged record of the deep mineralization driven by new production imported from above. By combining Apparent Oxygen Utilization (AOU) derived from dissolved O_2 distributions with time scales derived from transient tracers ^3H and ^3He JENKINS and co-workers arrived at estimates of New Production about an order of magnitude higher than assessed by other methods, roughly equalling rates commonly reported for Gross Primary Production (JENKINS, 1982; JENKINS & GOLDMAN, 1985; JENKINS, 1988). Reconciliation between this and various other approaches is of course being sought (e.g. PLATT & HARRISON, 1985; 1986; KEY, 1987). If nothing else these results have led us to have a fresh look at the concepts and techniques of Gross Primary and New Production. Between the tracer approach and other methods (for GPP and NP) there is definitely a difference with respect to sampling of the ocean. Otherwise the methods for assessment of GPP (e.g. ^{14}C bottle incubation) have again come under scrutiny, not surprisingly as possible artefacts have always been pointed out or suspected (STEEMAN-NIELSEN, 1952; GIESKES, KRAAY & BAARS, 1979; FITZWATER ET AL., 1982; ALLDREDGE and COX, 1982; WARD, 1984). In the PRPOOS experiment various incubation methods, including the novel $H_2{}^{18}O$ method (BENDER & GRANDE, 1987), did however give rates in general good agreement, provided that state-of-the-art techniques (e.g. ultraclean as to avoid trace metal interferences) are employed (GRANDE ETAL, 1988). Obviously further intercalibration studies, refinements as well as new methods for shipboard assessment of GPP and NP are continuously being sought. Remains the problem of inadequate in-situ sampling, never to be solved by shipboard observations alone. Both the tracer approach as illustrated above, as well as long term 'time series' (moorings, free drifting buoys, weather ships) would somewhat 'fill the

gaps' by providing time-averaged information. Otherwise only satellite observations (see below) would provide some synoptical sampling, albeit only of the sea surface (PLATT & SATYENDRANATH, 1988).

In many open ocean areas the surface waters are extremely oligotrophic most of the year. The nanomolar levels of limiting N species (nitrate, nitrite) can now be measured with special techniques (GARSIDE, 1982, 1985). This has revolutionized our thinking about the oligotrophic euphotic zone (PLATT & SATHYENDRANATH, 1988). Biological growth, or at least New Production, is virtually nil, except when during storm events new nitrate is mixed up along with richer deep waters (EPPLEY & RENGER, 1988; JENKINS, 1988). As a first (steady state) approximation the annual mean upward diffusive N flux would be balanced by the export through POM and DOM. The key role of now detectable ultralow N levels for primary and new production within the oligotrophic system deserves special attention, both directly in the field as well as in (laboratory or shipboard) incubation studies.

VERTICAL TRANSPORT OF PARTICULATE ORGANIC MATTER

The settling particles hypothesis (McCAVE, 1975) has been predominant in the past decade, and is supported by extensive evidence from sediment traps (HONJO, 1978, 1980; WAKEHAM ET AL., 1980; FOWLER & KNAUER, 1986; MARTIN ET AL., 1987; TAMBIEV, 1987). Thus far there has been a great deal of attention for the fecal pellet component, emphasizing the important role of the larger zooplankton in the export of particulate organic matter. However fractions other than fecal pellets should not be overlooked either (PILSKALN & HONJO, 1987). Compilations of VERTEX data suggest that greater than 75 % of the particle flux is recycled within the upper 500m (MARTIN ET AL., 1987; PACE ET AL., 1987), while recognizing the tremendous basin-wide variability of the C flux. In the deep ocean vertical gradients of the fluxes of organic matter (C/N/P, organic compound classes, etc.) point at significant microbial mineralization and transformation during settling (DE BAAR ET AL., 1983b; LEE & CRONIN, 1984; WAKEHAM ET AL., 1984; WATSON & WHITFIELD, 1985). From the observed seasonality of long term records (DEUSER & ROSS, 1980; BACON ET AL., 1985; DEUSER, 1986) we now also know that the settling velocity can be very rapid indeed, effectively bringing the C down to the seafloor within about one month after a plankton bloom. The rate of interactions between the larger

rapidly settling particles (e.g. fecal pellets) and the fine suspended matter can be estimated from U-Th series disequilibrium studies, focusing either on the deep ocean or the upper ocean box (BACON & ANDERSON, 1982; BACON ET AL., 1985; COALE & BRULAND, 1987). The gap between the finest suspended and larger, rapidly settling, particles is bridged by the formation, alteration and destruction of larger aggregates (ALLDREDGE, 1986) and the flux and sinking speed of such 'marine snow' can now also be measured (ASPER, 1987). Here it should be emphasized that insight gained from fluxes driven by one given ecosystem (e.g. North Pacific) do not necessarily apply to another (e.g. North Atlantic). For particle flux studies in JGOFS one would wish to perform both field experiments (process studies) and long term time series deployments in a limited number of judiciously selected ecosystems, for example the Sargasso Sea (DEUSER, 1986), central North Pacific, Weddell Sea (FISCHER ET AL., 1988) or Northwest Indian Ocean.

DISSOLVED ORGANIC MATTER

The alternative DOC regeneration model in its ultimate form would envision a parcel of water as a closed system: while travelling laterally and aging in the deep abyss its DOC is slowly converted to DIC by microbial decay, at the expense of dissolved O_2 that is being consumed. Recently published results for DOC, based on a careful modern rejuvenation of the High Temperature Combustion (HTCO) technique (SUGIMURA & SUZUKI, 1988) would appear to support exactly this hypothesis. The Apparent Oxygen Utilization (AOU) in the deep ocean would be virtually completely accounted for by mineralization of DOC. These results are further supported by an inverse relation between Dissolved Organic and Inorganic Nitrogen (DON, DIN) where DON results are also based on a HTCO method (SUZUKI, SUGIMURI & ITAH, 1985). The suggested oceanographically consistent relations (AOU/DOC and DON/DIN) are not unattractive and would ultimately render fluxes of settling particles to be insignificant. Another implication would be the significance of free ranging bacteria, rather than those attached on settling particles (KARL ET AL., 1988; CHO & AZAM, 1988). However both these relations as well as the underlying HTCO methodology are currently the subject of considerable debate (WILLIAMS & DRUFFEL, 1988; TOGGWEILER, 1988) and investigations (FITZWATER & MARTIN, BREWER & PELTZER, MANTOURA & PRESTON, all in progress).

One should finally bear in mind that the various sampling methods and ensuing particle size fractionations for marine aggregrates, settling particles, DOC, POC, (microbial) biomass, productivity measurements, etc. etc. more often than not are fundamentally different, providing an outlook through very different windows into the deep ocean: an appreciable portion of all the perceived discrepancies stems from the observer.

THE MICROBIAL LOOP

The omnipresence and abundance of bacteria and small protozoans in the marine ecosystem has in the past decade received considerable attention, not in the least because of the voracious predation of bacteria by protozoa (WILLIAMS, 1981, 1984; AZAM ET AL., 1983; HOBBIE & WILLIAMS, 1984). Relationship between bacterial biomass and chlorophyll (BIRD & KALFF, 1983) or primary productivity have been established within a range of marine systems. Most of the bacteria appear to be free living (rather than attached) utilizing Dissolved Organic Carbon (DOC) rather than particles as a food source (HODSON & AZAM, 1977; WILLIAMS, 1981). Through such studies the concept of a substantial microbial foodweb has been advanced, where the bacteria are sustained by a flow of dissolved organic matter derived from phytoplankton and herbivores. The dissipation of C through this passway has now also been demonstrated in a large scale field study (DUCKLOW ET AL., 1986). Within the euphotic zone the overall effect of the bacteria may be net remineralization (PLATT, 1985), enhancing the efficiency by retaining essential nutrient within the surface ocean (FROST, 1984, FASHAM, 1985). The conventional description of flows of energy and matter through trophic levels (phytoplankton, herbivores, carnivores, settling flux of fecal pellets) is now augmented by microbial pathways. For the deep ocean CHO & AZAM (1988) and KARL ET AL (1988) have now also provided evidence that free ranging bacteria play a major role in mineralization within the ocean interior as well.

Above insights provided by the microbiologists somewhat parallel the shift in thinking towards the cycling of Dissolved Organic Matter as a means for describing C fluxes in the sea, at the expense of the particle flux concept. Earlier we found substantial evidence for both the DOM paradigm and the (settling) POM concept. When taking into account the variety of pelagic ecosystems in different oceans as well as the current extreme undersampling of such ecosystems one would predict that

both mechanisms will finally prove to be of significance for the marine C cycle.

TRACE ELEMENTS

The cycling of C by the biota in the ocean is accompanied by the intentional or incidental entrainment of many trace elements (BRULAND, 1983; WONG ET AL., 1983). Vertical profiles of dissolved Cd, Zn, Ni, Cu (BOYLE, ET AL., 1976; 1977; SCLATER ET AL, 1976; BRULAND, 1980), Pd (LEE, 1983), Ba (CHAN ET AL., 1977), the Rare Earths (ELDERFIELD & GREAVES, 1982; DE BAAR ET AL., 1983a, 1985), Al (HYDES ET AL., 1988) as well as lateral sections (KREMLING and POHL, 1988) show significant correlations with the nutrients. It is unlikely though that all these metals would also act as truly biolimiting nutrients, more complex mechanisms (e.g. adsorptive scavenging) for entrainment with particle fluxes have to be proposed (BALISTRIERI ET AL., 1981; HONEYMAN ET AL., 1988). For example for many metals the settling flux is proportional to the overall mass flux (JICKELLS ET AL., 1984). The scavenging agent on the settling particles may well be some ferromanganese oxide coating (COWEN & BRULAND, 1985), again releasing entrained metals (e.g. Rare Earths, DE BAAR ET AL., 1988; GERMAN ET AL., 1987; SCHIJF ET AL., 1988) when dissolving under suboxic or anoxic conditions. In another instance (Ba) micro-environments are invoked (BISHOP, 1988).

Still, many transition metals like V, Cr, Mn, Fe, Co, Ni, Cu, Zn and Mo are indeed essential nutrients with known functionality in e.g. enzyme systems. However the oceanic distribution of most of the latter elements is not nutrient-like but apparently dominated by other processes such as overriding input functions (atmospheric, riverine, hydrothermal) or oxidation-reduction (photo)chemistry (LANDING & BRULAND, 1987; SAAGER & DE BAAR, 1989; SUNDA & HUNTSMAN, 1988). In some instances their availability may act as a significant limitation on rates of C-fixation, with Fe and Mn as examples (ANDERSON & MOREL, 1982; BRAND ET AL., 1983; MARTIN & FITZWATER, 1987; MARTIN & GORDON, 1988; COALE & BRULAND, 1988; HARRISON & MOREL, 1986). In the case of Cu the cupric ion activity primarily acts as toxicant, inhibiting rather than stimulating photosynthetic C-fixation (SUNDA & GUILLARD, 1976; ANDERSON & MOREL, 1978; BRAND ET AL., 1986), also through competition with Mn (SUNDA & HUNTSMAN, 1983). Likewise one finds Cd to act solely as a toxicant (BRAND ET AL., 1986), an apparent paradox when contemplating its highly significant correlation with oceanic phosphate. The

unraveling of the interaction of these trace elements with the biota has only just begun but is again crucial for assessing fluxes in the C cycle. Indirectly these metals may also, through shipboard contamination, affect the determination of primary productivity and clean ^{14}C-incubation techniques are required (FITZWATER ET AL., 1982). Several of these trace elements (Cd, Cu, Ni, Pb, Zn) are also discharged as man's wastes into the ocean and feedback loops to be resolved will undoubtedly be found to affect the oceanic C cycle, hence the atmospheric CO_2 level and the global heat budget.

GLOBAL SURVEY BY REMOTE SENSING

Remote sensing of the surface ocean with color scanners mounted on a satellite in principle allows worldwide coverage of pigment (chlorophyll) abundance at the sea surface (BROWN ET AL., 1985; ESAIAS ET AL., 1986). Corrections for atmospheric interferences have been well developed now for resolving the signals from the past overflights (1979-1986) of the Coastal Zone Color Scanner (GORDON and MOREL, 1983). The resulting data describes the surface abundance of chlorophyll pigments, the biochemical sites for C-fixation. From these distributions alone we have already gained considerable insight into the dynamics of the pelagic ecosystem (BANSE & MCCLAIN, 1986; ABBOTT & ZION, 1987; YODER ET AL., 1987; WALSH ET AL, 1987). In order to further pursue the desired global quantification of marine photosynthetic C-fixation (primary productivity) the satellite images need to be accompanied by proper algorithms. Development of such algorithms depends on simultaneous shipboard observations of pigments abundance (chlorophyll a) and primary productivity as 'ground truth' data for calibrating the numerical models (PLATT & SATHYENDRANATH, 1988; LOHRENZ ET AL., 1988). For more sophisticated assessment of various pigments one would routinely utilize HPLC methodology. Apart from the remote sensing application this technique of course also provides insight into composition and physical state of the plankton community, notably with respect to the important nanoplankton component (GIESKES ET AL, 1986, 1988). The above assessments of productivity are among the first and there is obviously room for further improvements in both precision and accuracy. Exactly such quantification with small overall error is one of the major objectives of the JGOFS program.

In 1991 new data will become available from the SeaWifs ocean color sensor scheduled to be flown on the Landsat-6 spacecraft (PUTNAM, 1987). Application of atmospheric corrections and algorithms for productivity would rely strongly on the expertise now available from the past CZCS experiment. Through coordination with carefully planned shipboard work (both local process studies as well as global survey lines) one should be able to produce in virtual real-time large scale maps of ocean color as well as productivity.

PUMPING CARBON INTO THE OCEANS

So far we dealt with the dissolution of CO_2 in seawater and its uptake into particulate and dissolved organic matter. Coccolithophorideae, planktonic foraminifera and planktonic ostracods do however also extract CO_2 (about 3 GtC per year) from surface seawater in the form of $CaCO_3$ shells. Most of these shells settle to the seafloor and thus constitute another flux of C into the deep sea (HONJO, 1980). Some of this $CaCO_3$ is being replenished by river input (KEIR & BERGER, 1985), the majority has to be regenerated within the oceans (SUNDQUIST, 1985) as to retain steady state (the net flux of Ca from hydrothermal systems being ignored; THOMPSON, 1983). Both the $CaCO_3$ preservation record in marine sediments (lysocline depth, CCD, etc.; KENNETT, 1982) and physical-chemical studies (BROECKER & PENG, 1982) agree on the fact that such dissolution of calcite shells takes place only at great depth, where the lower carbonate ion concentration allows undersaturation of the ambient seawater. However recent preliminary results suggest that an appreciable portion (as high as 50 %) of the calcite is in fact remineralized in the upper water column. From comparison of size fractions of living and dead (settling) foraminifera it appears that some 80 % of the individuals (mostly juveniles, BRUMMER ET AL, 1986) or about 30 % of the calcite never reaches the deep sea (BRUMMER, 1988; see also BERGER, 1976, his Figure 29.6). Dissolution in the upper ocean is presumably achieved within the shell, where a microenvironment (e.g. ALLDREDGE & COHEN, 1987) is invoked in which through microbial oxidation of organic matter the pH and carbonate ion level are lowered as compared to ambient seawater. More insight in this latter scenario as compared to the classical settling & thermodynamical dissolution model needs to be gained for a realistic assessment of $CaCO_3$ fluxes settling out of the euphotic zone.

In summary we now have four, rather than three (VOLK & HOFFERT, 1985) different routes for 'pumping' C into the deep ocean:

1. Dissolved bicarbonate carried down in deep water formation.
2. Dissolved Organic Carbon carried down in deep water formation.
3. Vertical flux of organic matter in large settling particles.
4. Vertical flux (sedimentation) of calcite shells.

For proper study of these four pumps it is crucial to also gain considerable insight into the interaction with many other chemical elements (N, S, P, Mn, Fe were mentioned as examples), notably with respect to ocean/atmosphere exchange, primary productivity and new production.

MODELING

Modeling has become an essential tool for both setting up the hypotheses before as well as interpreting the gathered data after the (field) experiment. In a rather extreme stance we might call the whole JGOFS excercise an effort towards developing a dynamic global model of the flow of C through the marine system. More modestly and realistically one may think of: models describing the nutrient field in the euphotic zone as a function of physical forcing (GLOVER & BREWER, 1988); foodweb models (FASHAM, 1985); U-Th decay series modeling as to provide time clocks for oceanic processes (e.g. BACON ET AL., 1985). Within the context of this short paper we cannot dwell extensively on the many aspects of marine modeling; excellent overviews are available FASHAM (1984), FRANSZ ET AL (1989), also specifically for an Ocean Flux Study (U.S. GOFS 4, 1987).

For example in a process study like the North Atlantic Pilot Study (see below) the field data collected will be used by an international modeling working group (SCOR/JGOFS W.G.) to implement and calibrate carbon flux models. Adequate models would describe quantitatively the rates of change of carbon content in biotic and abiotic ecosystem components, and transport processes in vertical and horizontal directions. Numerical integration may be applied to simulate seasonal variations, and to demonstrate the fate of assimilated carbon. This will also provide quantitative insight into annual budgets and balances.

TOWARDS AN INTERNATIONAL JOINT GLOBAL OCEAN FLUX STUDY

From above and other considerations has grown the consensus that quantification of the marine biogeochemical fluxes deserves a long term, international, research effort. After an initial Global Ocean Flux Study workshop at the National Academy of Sciences (Woods Hole, 1984) several more national workshops were held in for example the USA (Woods Hole, 1985; Princeton, 1985; Horn Point, 1986; Seattle, 1986), the United Kingdom (Royal Society, London, 1986; Bristol, 1986) as well as The Netherlands (Royal Netherlands Academy of Sciences, Amsterdam, 1986, 1987). Various Ocean Flux symposia, for example held during the January 1986 Ocean Sciences Conference in New Orleans (American Geophysical Union / American Society for Limnology and Oceanography) further served as an avenue for planning and informal contacts. Largely through these workshops and ensuing documents (see Reference section) national JGOFS related programs are being planned (Canada, China, Japan) or have already been launched (UK, FRG, France, The Netherlands, USA).

Joint Global Ocean Flux Study. - The international organization for the JGOFS program was firmly established at a meeting in February 1987 sponsored by the Scientific Committee on Oceanographic Research (SCOR) and held at Paris headquarters of its parent organization, the International Council of Scientific Unions (ICSU). Representatives from the FRG, France, UK, USA, Japan, Canada, India, China and The Netherlands attended and agreed on the main objectives for the JGOFS program (see below). Furthermore the development of a North Atlantic Pilot Program for 1989, 1990 among the adjoining countries (FRG, France, UK, Canada, USA, Netherlands) was recommended. Links with other global programs (WOCE, TOGA, IGBP) and agencies (IOC, CCCO, SCOR) at both international (Figure 8) and national level were also recognized cq. recommended. In October 1987 the SCOR established a 'Committee for JGOFS' with the task of *Planning and Implementation* of a program for the 1989-1999 period. Membership from the various participating countries is listed in Table 1. In January 1988 the Committee met for the first time in Miami (SCOR/JGOFS, 1988a); attention was paid to long term planning (1989-1999) as well as the immediate needs of the initial North Atlantic Pilot Study. The second meeting of the Committee and its guests was held in The Hague in September 1988 (SCOR/JGOFS, 1988b) in conjunction with a planning session of the Pilot Study group and the first meeting of the CCCO/JGOFS CO_2 panel.

SPECIFIC OBJECTIVES OF JGOFS

Having considered the national programmes the February 1987 Paris meeting agreed on a JGOFS program for the 1989-1999 decade whose main goal should be:

> *"To determine and understand on a global scale the processes controlling the time-varying fluxes of carbon and associated biogenic elements in the ocean, and to evaluate the related exchanges with the atmosphere, the sea floor and continental boundaries."*

This is the major objective of JGOFS and each of the individual research projects within the various national programmes will always be directly related to this central theme. The underlying rationale can be summarized as follows:

1. The need for fundamental field data:

Climate and the oceanic C cycle are linked; quantitative insight into the C cycle of the modern ocean is required as a benchmark for climate studies, with an emphasis on todays variability in time (from day to decade) and space.

The marine C budget is intimately related with that of other common (N, S, P) and uncommon (Fe, Mn, Cd, Cu, Th) chemical elements. Cycles of such biogenic elements deserve better understanding, also with respect to ocean/atmosphere exchange.

The existing archive database for marine primary productivity is biased in time/space and also suffers from artefacts of methods as they were used. Significant global coverage suitable for global budget modeling can never be realized from shipboard measurements.

Process studies, utilizing state of the art methodology, in a number of judiciously chosen areas would on the other hand be feasible and provide the basis for worldwide extrapolation based on satellite observation.

The important concept of new production and the ensuing F-ratio (new/primary production) needs to be defined more precisely through innovative experimental approaches in the field. Quantification in various oceanic areas is required as to obtain an assessment of the 'biological pump' of C leaving the euphotic zone.

Process studies of fluxes (primary productivity, new production) are to be complemented by species level studies of bacterio/phyto/zoo

plankton, their physiology, biochemistry, pigment characteristics and their interactions and significance within the overall marine foodweb.

Similarly the structure of the bentic community as related to regenerative fluxes from the seafloor deserves attention.

Organic chemical cycles in the ocean are relatively well understood and shown to be fascinating in themselves for several compound classes (FARRINGTON, 1987), but largely obscure for other compounds or the total organic pools (notably DOC, DON, DOP). Development of improved analytical methodology (e.g. MOPPER, 1986; SUGIMURA and SUZUKI, 1988) is crucial.

Internally consistent datasets encompassing physical, biological and chemical variables are needed for well defined areas, also for those parameters which can already be measured satisfactorily (e.g. salt, nutrients), as to constrain local, basinwide or global (mass balance) models.

2. The need for time series observations over several decades, as to allow observation of global change.

Time series of vertical fluxes collected with sediment traps, as already in place near Bermuda, at various key oceanic stations.

Time series of chemical and biological parameters, data to be collected from ships or buoys.

Time series of plankton species abundance and provenance, such as the UK Continuous Plankton Recorder Survey.

Time series of phytoplankton pigment abundance, sea surface temperature and other parameters as available from satellite sensors.

3. Opportunities from new technology:

Satellite observation (NOAA/AVHRR) already provides data in the Infrared (Sea Surface Temperature) and Visible (Total Particulate Matter) region. Assessments of wind stress, sea state and ocean currents will become available from ERS-1 and TOPEX/POSEIDON. Most essential will be the ocean color data from the compact wide field sensor (Sea-WIFS) which will be launched aboard U.S. LANDSAT-6 in late 1991. The selected wavelengths of Sea-WIFS are compatible with those of CZCS (1979-1986) such that the now available algorithms for processing CZCS data can be utilized.

Modern computers have led to the now available sophisticated models of the ocean, models quite suitable for biogeochemical purposes, where validation of course requires field data.

New or currently developed methodology is suitable for the necessary application at a larger scale: realistic measurements of nanomolar levels of nitrate, nitrite, ammonia in oligotrophic waters; accurate CO_2 determination by coulometry; clean techniques for assessing primary productivity; plankton characterization by flow cytometry; in situ sensors (pH, CO_2); photosynthetic pigments by HPLC; organic biomarkers and their transformations; DOC/DON; sediment traps and bentic instrument packages.

Large scale physical programmes like WOCE (1986, 1988) AND TOGA are also oriented towards assessing the role of the ocean in climate; the physical insights (circulation, heat and salt exchange, predictive models, etc) to be gained with state of the art physical tools will complement the JGOFS focus on marine biological and chemical driving forces of climate.

When translating these major objectives into a JGOFS program for 1989-1999 it is obvious that the Planning and Implementation of the (field) work can be divided into roughly three categories:

1. Process Studies
2. Time Series
3. Global Survey

PLANNING: 1989-1999 JGOFS

Effective short term planning has led to the ongoing North Atlantic Pilot Study in 1989-1990 (see below), an important activity both scientifically as well as for gaining experience with such issues as data management, modeling, data quality and intercalibration, logistics. Long term planning for the overall JGOFS program throughout the nineties is currently being worked out by the SCOR Committee for JGOFS (SCOR/JGOFS, 1989). Next to the field studies (see below) several other aspects of the program are now being developed:

Modeling will play a major role at all stages of the program. Recognizing this prime importance a special Modeling Working Group has been established as the first of an otherwise very limited number of SCOR/JGOFS Working Groups (Table 2). This WG has the task of identifying and fostering a community of modelers, and developing effective two-way communication between modelers and observationalists. Liaisons with modeling groups in the IGBP and WCRP would also be desirable (ZEITSCHEL, 1988). Organization of an international modeling workshop is envisioned as a logical first step. Next to these long term commitments the WG will also play an active role with respect to the Pilot Study.

Data management is required for the overall program. The second, and thus far only other SCOR/JGOFS Working Group (Table 2), will address this issue, obviously also in relation to overall data quality and intercalibration. The WG will formulate a strategy to develop and distribute a historical archive of biogeochemical data, notably in situ and satellite pigment data. Liaisons with other global programs (WOCE, IGAC, IOC, TOGA) and oceanographic data centers also need to be established. With respect to the field studies timely data exchange methods between both the various teams (e.g. shipboard, remote sensing, modeling) and the participating national groups is required. Throughout the Pilot Study the WG will already play an active role, also with an eye towards gaining experience for the long term study.

Briefly the field oriented components will be addressed (SCOR/JGOFS, 1989) within the aforementioned categories: (1) process studies at specific sites, (2) time series and (3) global survey, where the option of basinwide studies as intermediate between (1) and (3) is furthermore conceivable.

1. Process Studies. - For several carefully selected sites representing the range and diversity of pelagic ecosystems integrated studies of several groups of processes and mechanisms will be developed. At each site a suite of observations will be made:

In the upper ocean the relationship between physical forcing, biogeochemical fluxes and the structure of the ecosystem needs characterization in terms of flows of energy and matter between biotic and abiotic components. Air/sea exchange of various components plays a major role and will be addressed in a joint effort with atmospheric scientists. Standing stocks and turnover rates (e.g. primary productivity, secondary production, bacterial biomass and turnover) of the various biological components need to be quantified, with intricate relations between compartments well understood. Geochemical variables to be assessed include dissolved nutrients (also at trace levels), pigments, (radio)tracers and dissolved gases. At each specific site the ecosystem structure and turnover rates will be related to remote sensing observations, also with an eye towards developing algorithms (e.g. PLATT & SATHYENDRANATH, 1988) to be applied in basin-wide studies and eventually for overall budget calculations linked to the global survey.

The relationship between the upper ocean and the fluxes of POM and DOM leaving the euphotic zone (i.e. new production) will be studied through observation of standing stocks and settling fluxes of particulate matter in various size classes at the bottom of the euphotic zone, through assessment of the nitrogen cycle with emphasis on nitrate fluxes through the thermocline, through quantification of fluxes driven by active vertical migration of marine organisms and through measurement of the amount, chemical composition and trunover rate of Dissolved Organic Matter. Free floating sediment traps, submersible pumps as well as diver operated samplers will provide insight into the particulate matter and its rates of aggregation and disintegration. Time constants for processes of particle formation, chemical scavenging of trace elements and rates of new production can be derived from natural radioisotopic decay series (e.g. ^{234}Th).

Transformations within the ocean interior occur at particles of all size classes with corresponding range of settling velocities, as well as in the dissolved phase. Biological processes like consumption and defecation by (larger) zooplankters or microbial conversion, as well as physical mechanisms (aggregation, disintegration) and chemical processes (desorption) need quantification, where the U-Th series disequilibria would again provide tools for establishing rates. Remineralization also

occurs in the dissolved phase (DOM) where the activity of free ranging bacteria also in relation to conversion of dissolved oxygen will be addressed. The overall standing stock of DOM and its organic chemical and elemental composition deserve special attention. Lateral transport terms of both particulate and dissolved components needs characterization in a 3-dimensional framework.

At the sediment-water interface the influx of settling matter is further mineralized by the benthic community, where dissolved regenerated material may again drive upward quasi-diffusive fluxes into the deep waters. The structure and activity of bentic communities needs to be understood and quantified, the geochemical fluxes downward and upward can be studied with tools like deep moored sediment traps or bentic chambers. The seasonal fluctuations of large size class material arriving at the seafloor will be investigated in relation to its chemical composition (e.g. labile components, N/C ratio) as a food source for the benthos. The upward return fluxes of remineralized constituents are driven by microbial turnover, by bioturbation as well as through interaction with mineralization of inorganic components such as biogenic silica and calcite. The residual modified flux is buried in the sediment and constitutes the sedimentary record which will be employed to assess temporal variation of the oceanic C cycle over geological time scales.

Selection of specific sites for these process studies is done through evaluation of suitable basins in which subsequently the exact site will be determined. The SCOR Committee for JGOFS currently investigates the merits of four different basins:

- The Pacific Basin as a province for a study site will be addressed at a SCOR/JGOFS sponsored workshop in Hawaii, 1989.
- The Antarctic or Southern Ocean is known to play a major role in the physical and biological fluxes of CO_2 between ocean and atmosphere. It also includes a major region for massive downwelling for which the effect on the dissolved fluxes of organic and inorganic C need further investigation. Between SCOR/JGOFS and established Antarctic research institutions (Australia, FRG, UK) the scientific and organizational aspects of a Flux Study in the Southern Ocean are currently investigated.
- The monsoonal driven seasonal upwelling and productivity regime in the (North West) Indian Ocean is quite remarkable and also

significant with respect to global C fluxes. An inventory of its merits as a study site has already been provided (SMITH, 1988).

- The North Atlantic Ocean has already been selected as the site for the Pilot Study, both for scientific and logistical reasons. Continuation of the North Atlantic Study beyond the Pilot stage is contemplated, where expansion to a basin-wide study would be one option.

2. Time Series. - Within global programs like JGOFS or IGBP there is the need to observe long term (climatic) trends through occupation of time series stations. For JGOFS such time series would also provide information on temporal variability over diel to annual time scales. In addition to continuation of already existing marine time series records (e.g. weather ships, sediment trap station off Bermuda (DEUSER, 1986), Continuous Plankton Recorder Survey) an extension is conceivable, for example long term sediment trap deployments in other areas (e.g. Biotrans station at 20°W, 47°N, or station near Hawaii, or at Ocean Station Papa). Next to such existing technologies options are to be explored for utilization of new technology, for example long term deployment of chemical/biological sensors on moored or drifting buoys.

3. Global Survey. - For determination of C fluxes on a truly global scale the only possible avenue is to measure by satellite, from conventional surface vessels one would never achieve synoptical global coverage. Large components of the JGOFS program are geared towards developing and calibrating the methodology to utilize satellite data in this manner. From 1991 onwards the ocean color data to be gathered with the Sea-WIFS scanner will be combined with 'ground-truth' pigment data from ships observations as to arrive at calibrated images of ocean color. Linkage with the process studies (1) will subsequently provide the additional algorithms required for deriving estimates of primary productivity. Close collaboration with the WOCE Hydrographic program would furthermore allow continuous shipboard recording of surface pigments along the extensive WOCE cruise tracks, data enhancing the accuracy and precision of worldwide estimates of pigment abundance and primary productivity. In addition to the ocean color data there already is valuable information (e.g. Sea Surface Temperature) being gathered from the NOAA/AVHRR system. Other variables collected by satellite such as sea state, surface height (altimetry) will also be applied, for example for assessment of rates of air/sea gas exchange.

IMPLEMENTATION: THE NORTH ATLANTIC PILOT STUDY

Following above objectives delegates from several countries (FRG, France, UK, Canada, USA, Iceland, Japan, China, The Netherlands) at the September 1987 planning meeting (Paris) discussed viable options for a Pilot Program in 1989, 1990 (SCOR, 1987b). Scientific rationale and logistics, as well as already existing or planned national programmes (e.g. BIOTRANS station at 48 N of FRG; 1988 APNAP time series study of The Netherlands) were taken into consideration. Six countries (FRG, France, UK, Canada, USA, Netherlands) agreed in principle on a coordinated study of a section along 20°W running from Iceland to Madeira, each country bringing in their own research vessel(s). This concept was further worked out, both within the individual member countries as well as during Pilot Study Planning Meetings in Plymouth (U.K.), The Hague and Monterey in April, September and December 1988 respectively (SCOR, 1988a,b). Overall coordination lies with a small team of national representatives (Table 3).

The Pilot Study will focus on three areas (33°N, 48°N, 60°N) along the 20°W transect (Figure 9), with a special emphasis on mesoscale variability. Throughout the seasons a wide range of mixing regimes will be encountered (Figure 10). This combined with the regime of light and temperature leads to the occurence of very distinct spring and fall plankton blooms. Through extensive coverage with research vessels (Figure 11) one would expect the 'bloom experiment' to be successful, not only in capturing the actual blooms, but also the important time frames leading towards and succeeding the blooms. Considerable effort will be dedicated towards measuring several critical variables simultaneously and on each ship. This way one aims at producing a data set of homogenous high quality (Table 3), suitable also for large scale modeling excercises. With respect to vertical fluxes of settling particles both free floating sediment traps (Figure 12) as well as deep moored trap arrays (Figure 13) will be deployed. Remote sensing observations will be available from the NOAA/AVHRR system and will be processed in most of the participating countries. In anticipation of the 1991 launch of the SeaWIFS satellite ocean color scanner the target areas will also be sampled during spring 1989 with the NASA Airborne Oceanographic LIDAR flown at about 100m altitude aboard a NASA P-3A aircraft (LEWIS, 1988).

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TABLES

B. Zeitschel (chair)	FRG	P.G. Brewer (vice-chair)	USA
H.J.W. de Baar	The Netherlands	M. Fasham	UK
O. Brown	USA	I. Koike	Japan
S.E. Calvert	Canada	K. Kremling	FRG
K.L. Denman	Canada	J.F. Minster	France
J.C. Duplessy	France	T. Platt	Canada
H. Elderfield	UK	D. Hu	China
R. Eppley	USA	S.B. Tambiev	USSR

Corresponding members:

J. McCarthy	IGBP Scientific Committee Geosphere-Biosphere Program
T. Nemoto	IGBP/SCGP
S. Krishnaswami	SCOR WG 71, Particulate Biogeochemical Processes
M.O. Andreae	SCOR WG 72, Ocean as Source/Sink for Atmospheric Constituents
C.S. Wong	SCOR WG 75, Methodologies Oceanic CO_2 Measurements
R.F.C. Mantoura	SCOR WG 78, Photosynthetic Pigments in Seawater
E. Sundquist	SCOR WG 79, Geological Variation CO_2 and C cycle
M. Whitfield	SCOR WG 80, Phase Transfer Processes Estuarine Trace Metals
A. McEwan	SCOR/IOC Joint Committee on Climatic Changes and the Ocean

Table 1. The SCOR Committee for JGOFS and its corresponding members.

SCOR/JGOFS Working Group on Modeling

M. Fasham	chair	UK
G. Fransz		Netherlands
T. Platt		Canada
G. Evans		Canada
J. Sarmiento		USA
G. Radach		FRG

SCOR/JGOFS Data Management Working Group:

T. Platt	chair	Canada
G. Flierl	USA	
M. Jones	UK	
G. White	Canada	
C. Steinen	FRG	

Coordination Group North Atlantic Pilot Study:

S. Piotrowicz	Overall coordinator (USA)

T. Platt	Canada
B. Zeitschel	FRG
H. de Baar	The Netherlands
M. Whitfield	UK
Dr. H. Ducklow	USA

Table 2. Working groups of the SCOR Committee for JGOFS and Coordination Group for the North Atlantic Pilot Study.

1. Meteorology and Positioning.
2. CTD, Probe for dissolved O_2, Fluorometry probe.
3. Dissolved O_2 by titration.
4. Dissolved nutrients
5. Optics
6. Dissolved CO_2 chemistry.
7. Particulate Organic Carbon and Nitrogen
8. Dissolved Organic Carbon
9. Pigments, chlorophyll
10. Biomass - Bacteria and Cyanobacteria
11. Biomass - Mesoplankton
12. Biomass - Microplankton
13. Primary production by ^{14}C
14. Primary production by O_2.
15. New production by ^{15}N
16. Bacterial production
17. Grazing - Mesoplankton
18. Grazing - Microplankton
19. Sediment traps - drifting
20. Sediment traps - deep, moored.

Table 3. Core measurements during the North Atlantic Pilot Study. In principle all measurements are executed by all participating ships/countries. Protocols are standardized with recommended precision and adequate intercalibration.

FIGURE CAPTIONS

Figure 1. The global Carbon budget, after MOORE & BOLIN (1986) and others. Inventories [1 Gigaton Carbon = 10^{15} gram C] and Fluxes [GtC/year] are large relative to Atmospheric CO2 content and its annual increase. Large arrows for natural fluxes, small arrow fluxes arise from human activity.

Figure 2. CO_2 concentrations ('best estimates') and smoothed values (spline function) in parts per million (volumetrically) plotted against age in the Vostok record (upper curve) and atmospheric temperature change derived from the isotopic profile (lower curve). The deuterium scale corresponds to values after correction for deuterium changes of ocean water. Taken from BARNOLA ET AL. (1987)

Figure 3. The increasing level of atmospheric CO_2 as a function of latitude over the 1980-1986 period. The pronounced seasonal oscillation in the Northern Hemisphere reflects the growth season of terrestrial plants, the latter being more abundant due to the predominance of landmasses in the Northern Hemisphere.

Figure 4. Trends in seasonal and annual mean temperature as measured from 1735 onwards, data from the Royal Netherlands Meteorological Institute (KNMI). Values shown represent means over ten year time intervals. The accuracy is 0.5 °C. The winter series is extended by inclusion of (indirect) data on frozen conditions of canals used by tow barges, taken from diaries of inland shipping companies. Calibration of the latter versus the first record over the 1735-1839 period yields an accuracy of 0.7°C for the canal record. The overall winter and spring records suggest a weak warming trend, not observed in the annual mean. Taken from VAN DEN DOOL ET AL. (1978).

Figure 5. One of the many 'ice-scapes' produced in the seventeenth century, at the peak of the Little Ice Age. For the purpose of reproduction we selected an etch rather than one of the many color paintings. "HYEMS" or Winter, one of a series representing the four seasons by Jan van de Velde, approx. 1617. The Latin text translates: 'The cold does not slow one down and deserves no resentment. Ovidius, you were wrong: look at our young lads; See how they step on frozen rivers, with skate-irons under their feet, and in long rows they swing through the pastures'. Taken from DE GROOT (1979). Todays skaters within The Netherlands definitely resent the prospect of global warming.

Figure 6. The partial pressure of CO_2 in surface seawater (1-15m) of the North Atlantic Ocean as measured during the TTO expeditions. A correction term for the surface excess of O_2 has been applied. This map does not represent synoptic sampling but a clear trend of over- cq. undersaturation in equatorial cq. subpolar waters is suggested. Taken from BREWER (1986).

Figure 7. Simplified scheme of compartments and fluxes for C in the upper ocean. The large arrows depict the four pathways for 'pumping' Carbon into the ocean interior.

Figure 8. International links of the JGOFS program.

Figure 9. The 20°W transect as studied in the North atlantic Pilot Study. The emphasis will be on the assessment of temporal and spatial scales at the three indicated stations: 30°N, 48°N and 60°N.

Figure 10. Mixed layer depth versus month at 30, 40, 47, 54, and 61 °N latitude along the 20°W Pilot Study transect. Original data taken from ROBINSON ET AL. (1979).

Figure 11. Seasonal coverage by participating research vessels during the North Atlantic Pilot Study along the 20°W transect, 1989.

Figure 12. Seasonal coverage with drifting sediment traps in or below the euphotic zone during the North Atlantic Pilot Study along the 20°W transect, 1989.

Figure 13. Seasonal coverage with deep sediment trap mooring arrays during the North Atlantic Pilot Study along the 20°W transect, 1989. Rotating sampling cup assemblies, sampling intervals of 2-4 weeks.

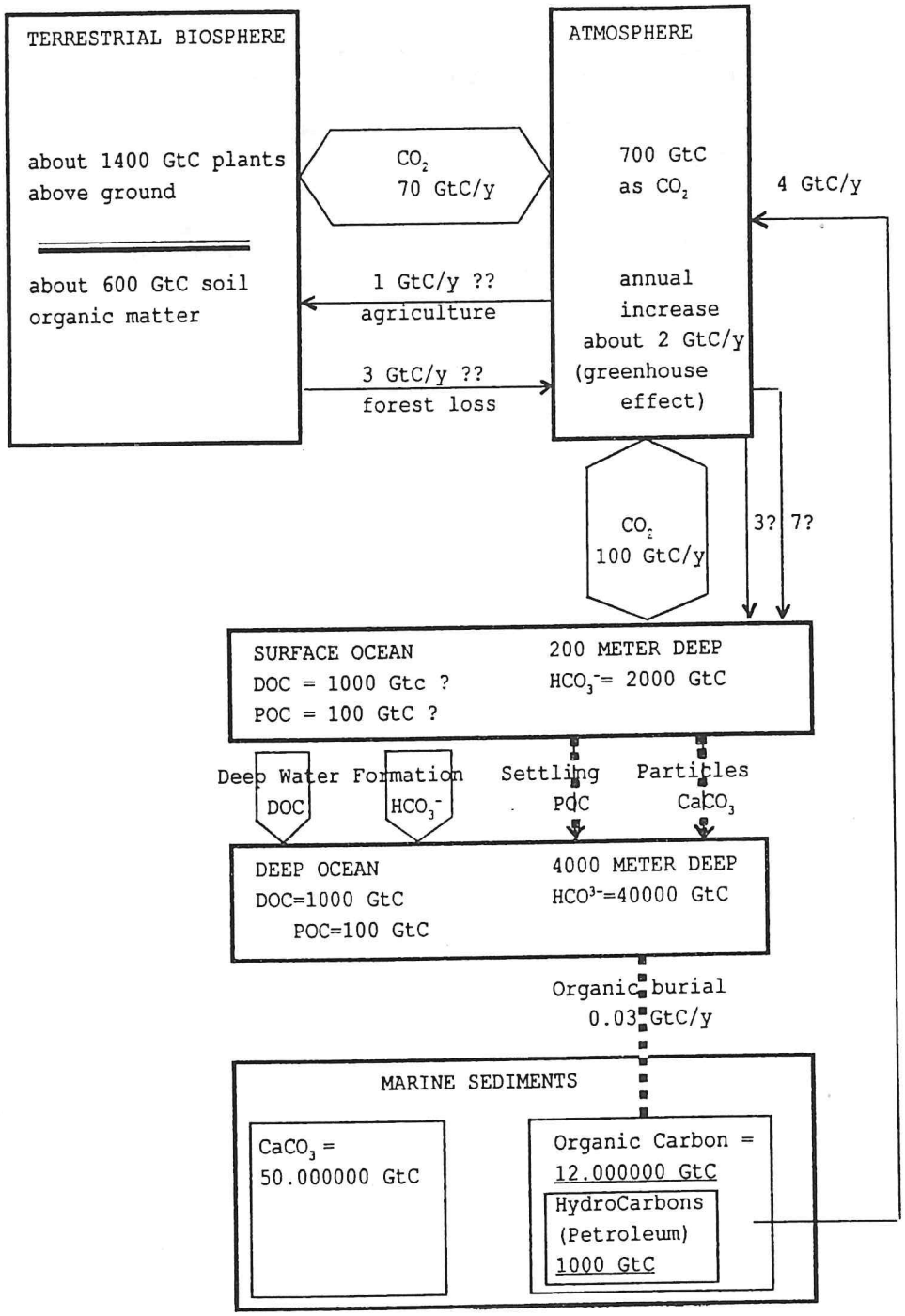

TERRESTRIAL BIOSPHERE
about 1400 GtC plants above ground
about 600 GtC soil organic matter
CO_2 70 GtC/y
1 GtC/y ?? agriculture
3 GtC/y ?? forest loss
ATMOSPHERE
700 GtC as CO_2
annual increase about 2 GtC/y (greenhouse effect)
4 GtC/y
CO_2 100 GtC/y
3? 7?
SURFACE OCEAN
DOC = 1000 Gtc ?
POC = 100 GtC ?
200 METER DEEP
HCO_3^-= 2000 GtC
Deep Water Formation
DOC
HCO_3^-
Settling
POC
Particles
$CaCO_3$
DEEP OCEAN
DOC=1000 GtC
POC=100 GtC
4000 METER DEEP
HCO^{3-}=40000 GtC
Organic burial 0.03 GtC/y
MARINE SEDIMENTS
$CaCO_3$ = 50.000000 GtC
Organic Carbon = 12.000000 GtC
HydroCarbons (Petroleum) 1000 GtC

Fig. 1

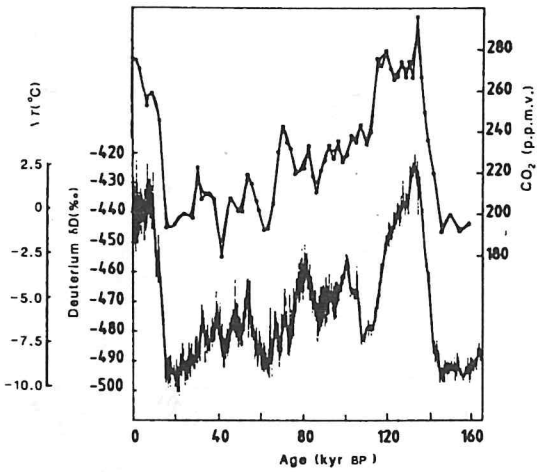


Fig. 2 CO_2 concentrations ('best estimates') and smoothed values (spline function) in p.p.m.v. plotted against age in the Vostok record (upper curves) and atmospheric temperature change derived[14] from the isotopic profile (lower curve). The deuterium scale corresponds to values after correction for deuterium changes of oceanic water[14].

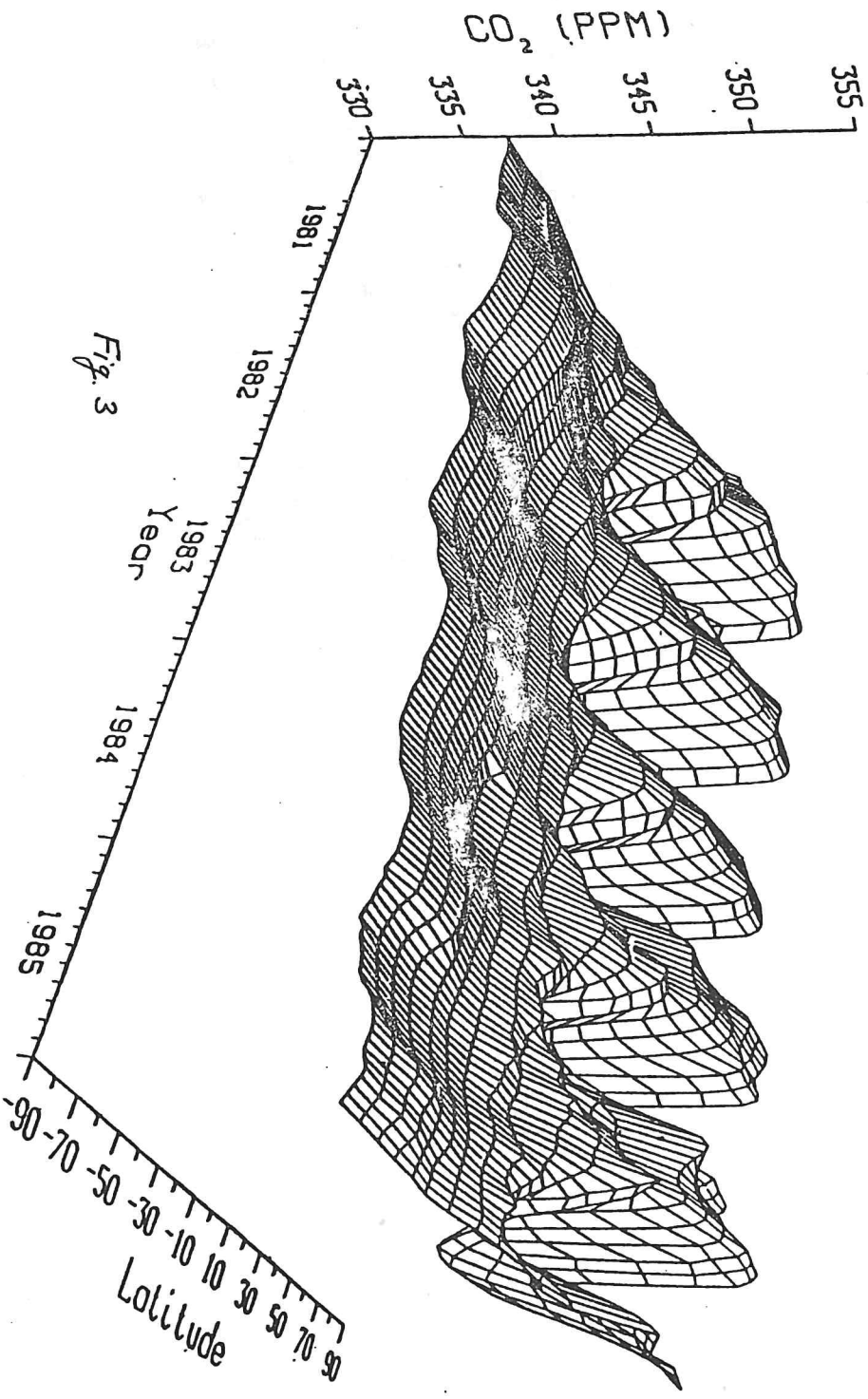

CO_2 (PPM)
330
335
340
345
350
355
1981
1982
1983
1984
1985
Year
-90 -70 -50 -30 -10 10 30 50 70 90
Latitude

Fig. 3

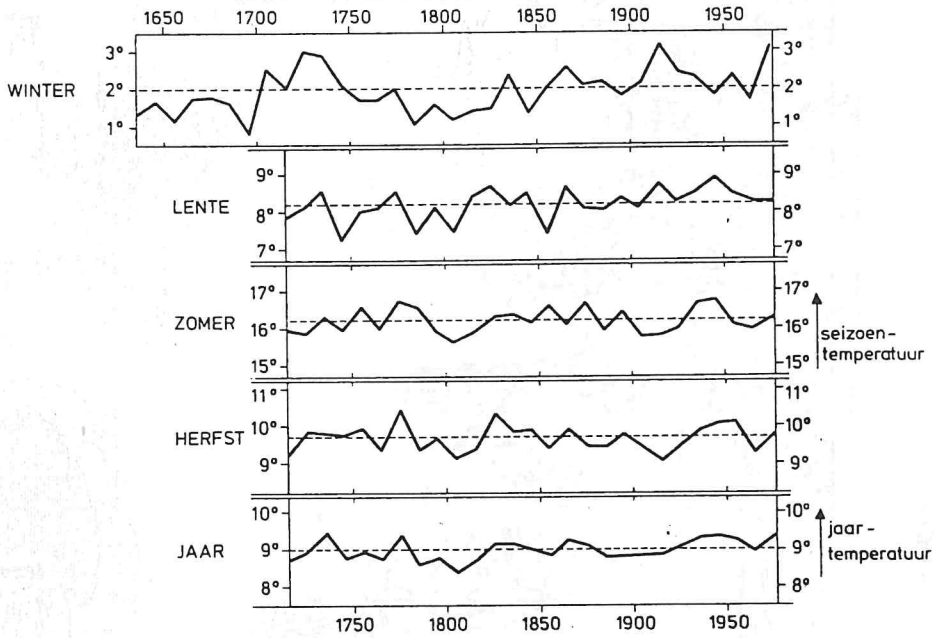

1650
1700
1750
1800
1850
1900
1950
WINTER
3°
2°
1°
LENTE
9°
8°
7°
ZOMER
17°
16°
15°
HERFST
11°
10°
9°
JAAR
10°
9°
8°
seizoen-
temperatuur
jaar-
temperatuur

Fig. 4

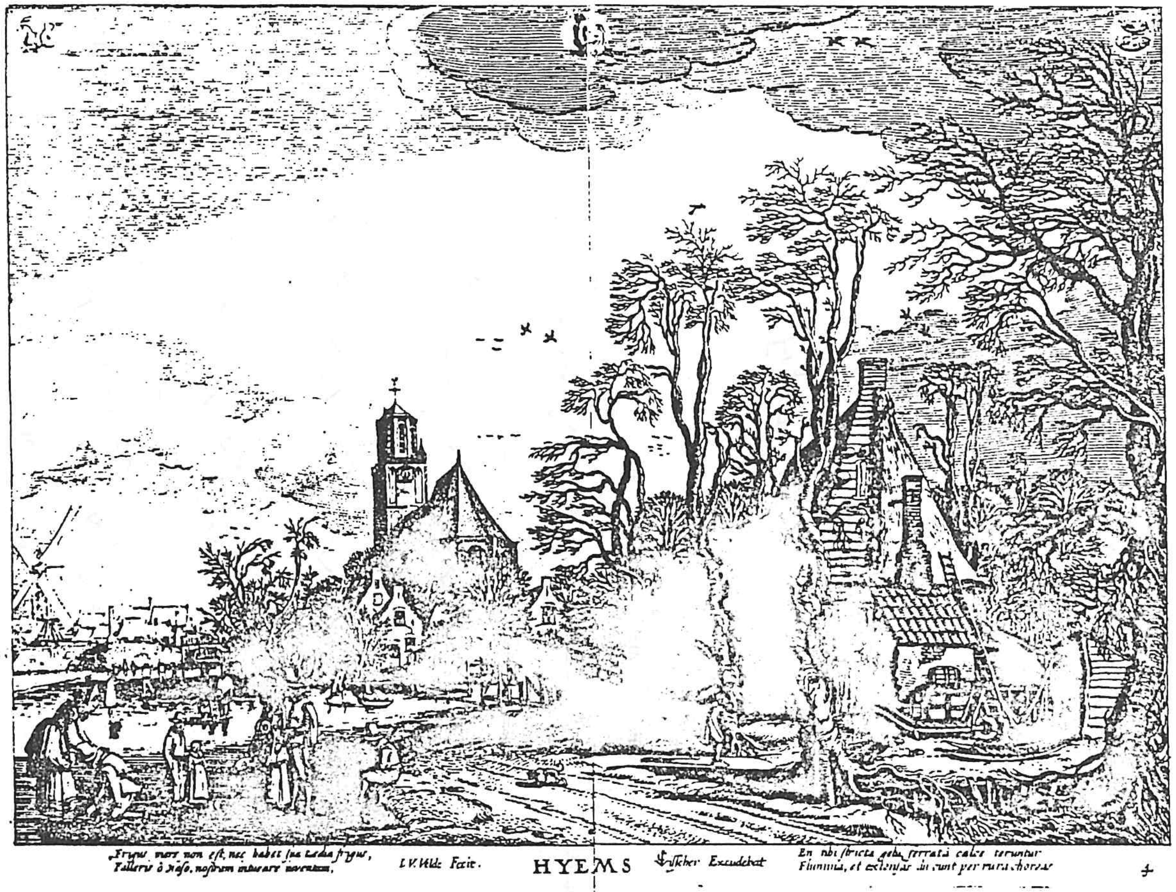


Fig. 5

TTO SURFACE (1-15M)

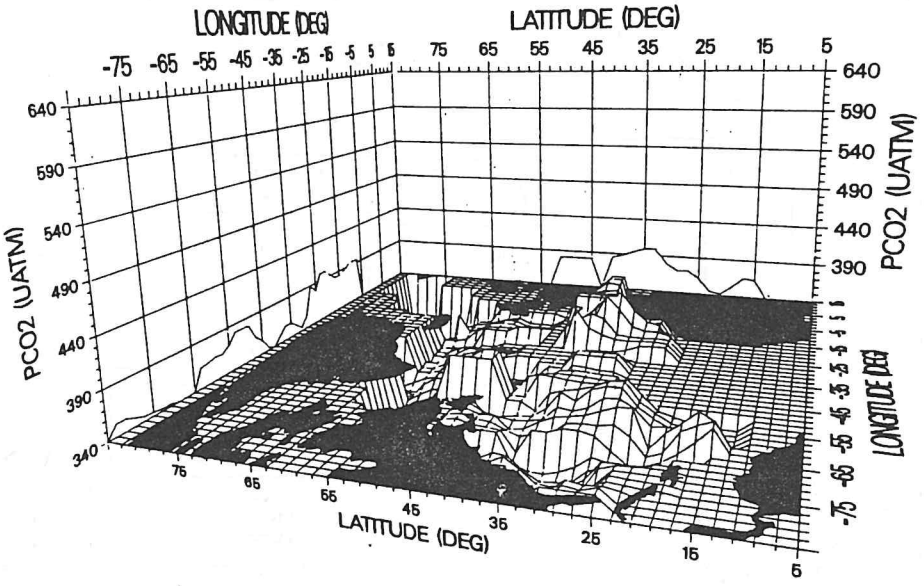
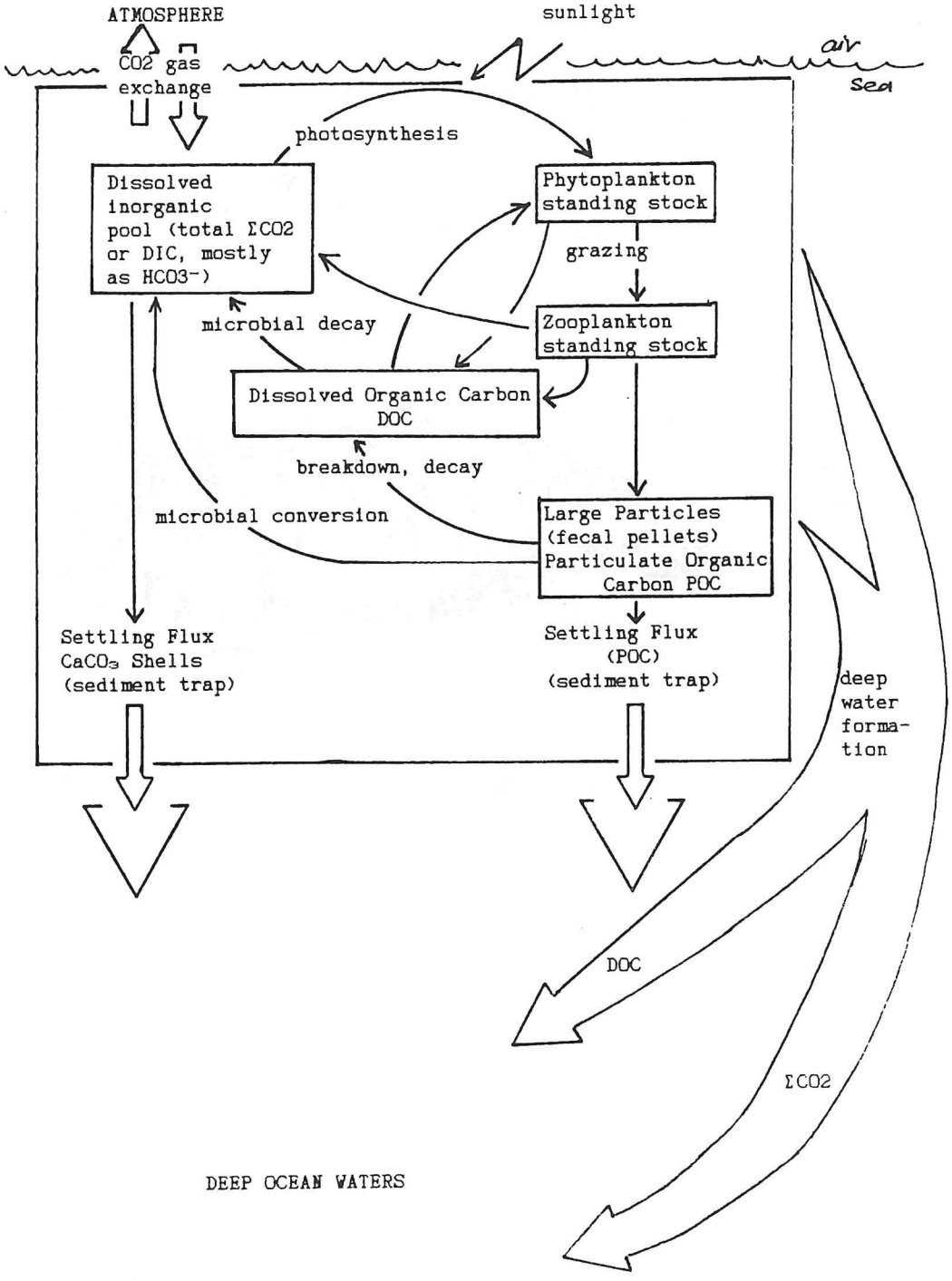


Fig. 6


ATMOSPHERE
sunlight
air
sea
CO2 gas exchange
photosynthesis
Dissolved inorganic pool (total ΣCO2 or DIC, mostly as HCO3⁻)
Phytoplankton standing stock
grazing
microbial decay
Zooplankton standing stock
Dissolved Organic Carbon DOC
breakdown, decay
microbial conversion
Large Particles (fecal pellets) Particulate Organic Carbon POC
Settling Flux CaCO3 Shells (sediment trap)
Settling Flux (POC) (sediment trap)
deep water formation
DOC
ΣCO2
DEEP OCEAN WATERS

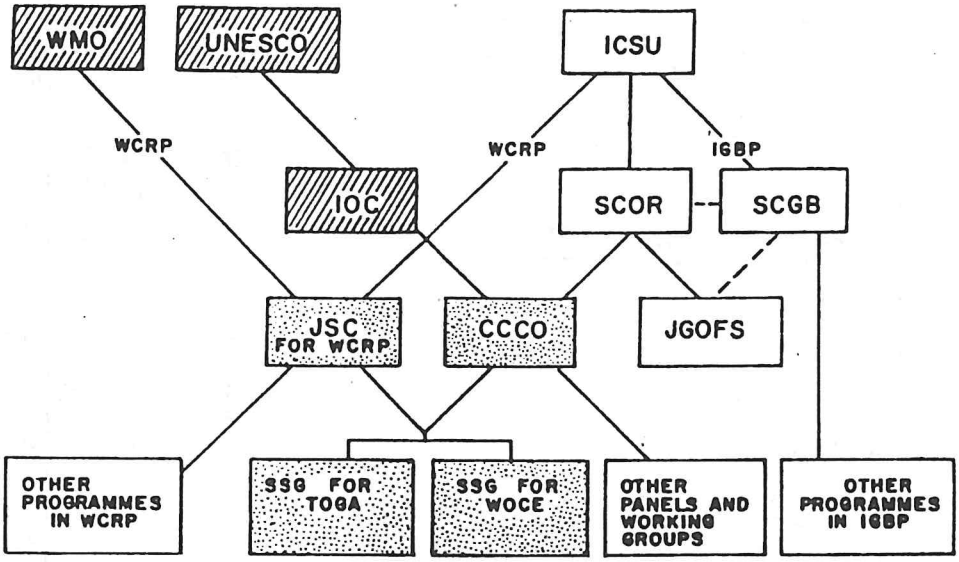

WMO
UNESCO
ICSU
WCRP
WCRP
IGBP
IOC
SCOR
SCGB
JSC FOR WCRP
CCCO
JGOFS
OTHER PROGRAMMES IN WCRP
SSG FOR TOGA
SSG FOR WOCE
OTHER PANELS AND WORKING GROUPS
OTHER PROGRAMMES IN IGBP
LEGEND:
FORMAL SPONSORSHIP
INFORMAL RELATIONSHIP
NON-GOVERNMENTAL ORGANIZATION
INTERGOVERNMENTAL ORGANIZATION
COSPONSORED BODY

Fig 8

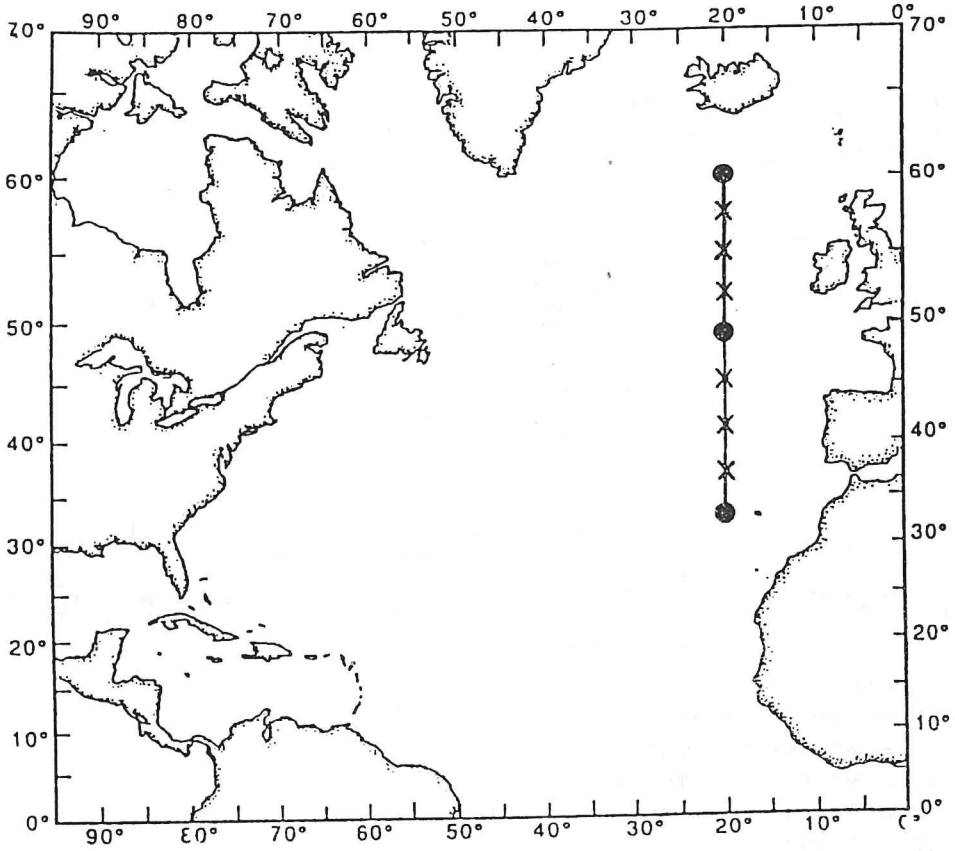


Fig. 9

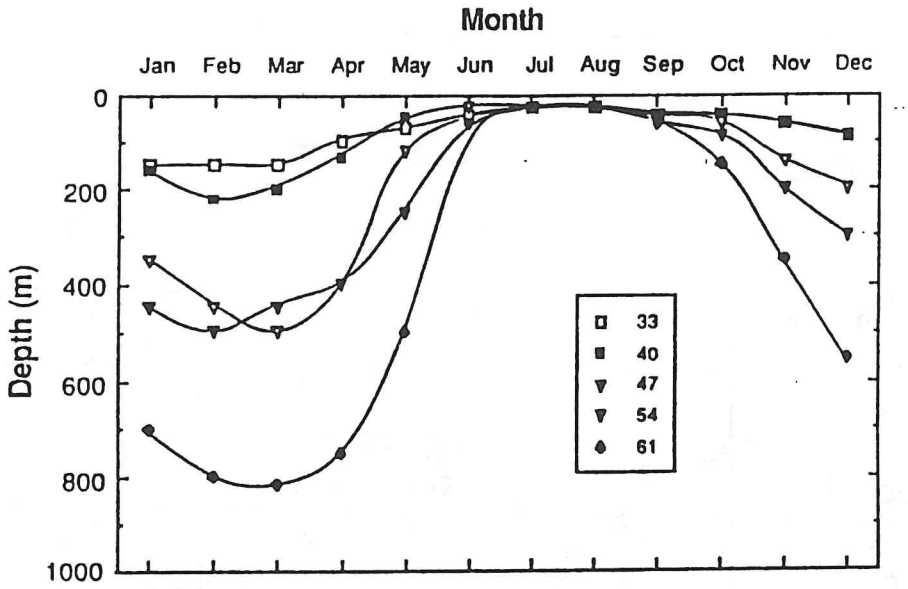
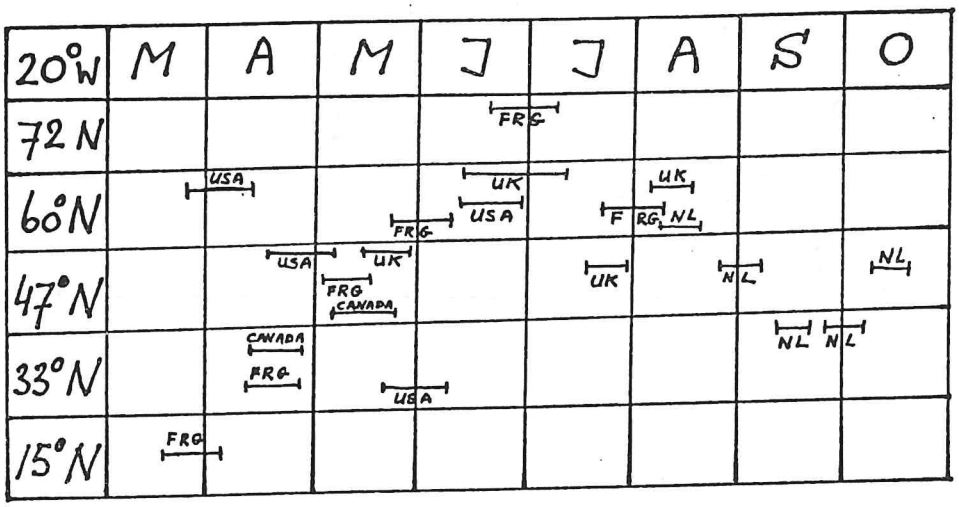

Month
Jan
Feb
Mar
Apr
May
Jun
Jul
Aug
Sep
Oct
Nov
Dec
Depth (m)
0
200
400
600
800
1000
33
40
47
54
61

Fig. 10

JGOFS North Atlantic Pilot Study 1989 - 1990

Seasonal Coverage at the three "Super Stations" (60/47/33)

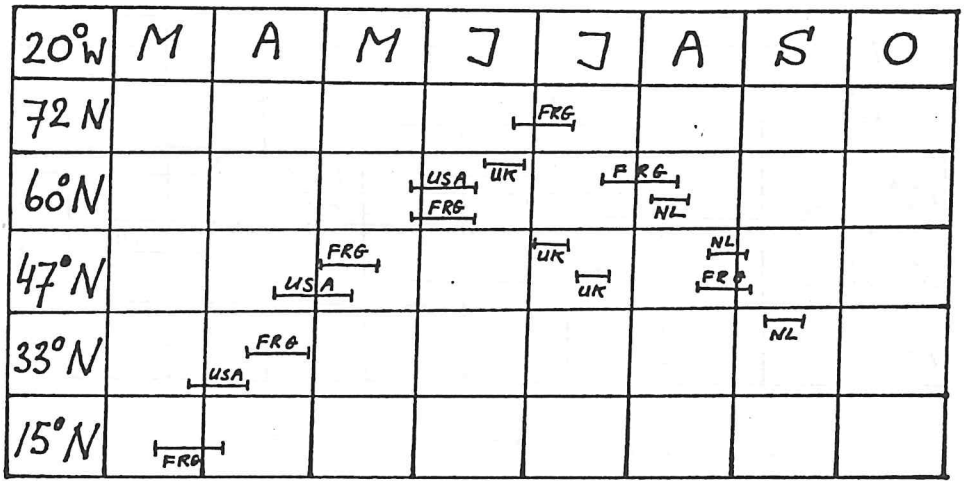


SURFACE VESSELS

Fig 11

JGOFS North Atlantic Pilot Study 1989 - 1990

Seasonal Coverage at the three "Super Stations" (60/47/33)

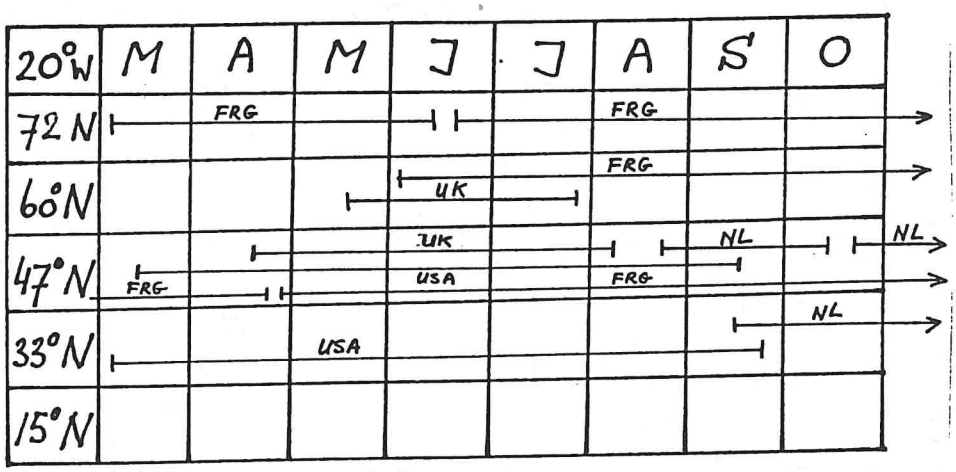


DRIFTING SEDIMENT TRAPS

Fig 12

JGOFS North Atlantic Pilot Study 1989 - 1990

Seasonal Coverage at the three "Super Stations" (60/47/33)



DEEP MOORED SEDIMENT TRAPS

Fig 13