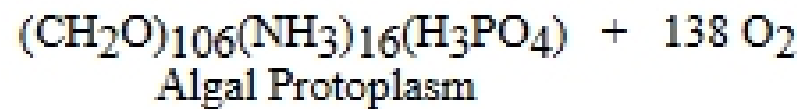


Respiration corresponds to the RKR reaction run backwards. This means that the dissolved oxygen concentration is a tracer for respiration. In the euphotic zone there is usually excess oxygen above atmospheric saturation because there is net biological production. Production of  $O_2$  by primary production is greater than consumption of  $O_2$  by respiration. Below the euphotic zone there is only consumption by respiration so the oxygen goes down as respiration proceeds. In some parts of the open ocean with restricted circulation all of the dissolved oxygen is consumed by respiration. Examples of these locations are the oxygen minimum zones in the eastern tropical north and south Pacific and Arabian Sea. At these locations respiration proceeds using nitrate to combust the organic matter. In some enclosed basins like the Black Sea, Cariaco Trench, Saanich Inlet and Framvaren Fjord respiration also uses up all the nitrate and sulfate is reduced to hydrogen sulfide ( $H_2S$ )

### Aerobic respiration

Oxygen is consumed and nutrients are released.



↓      bacteria



The oxidation of the  $NH_3$  in organic matter to  $NO_3$  is referred to as nitrification.

### Apparent Oxygen Utilization

We can calculate the extent of respiration by calculating a parameter called the Apparent Oxygen Utilization or AOU. AOU is defined as:

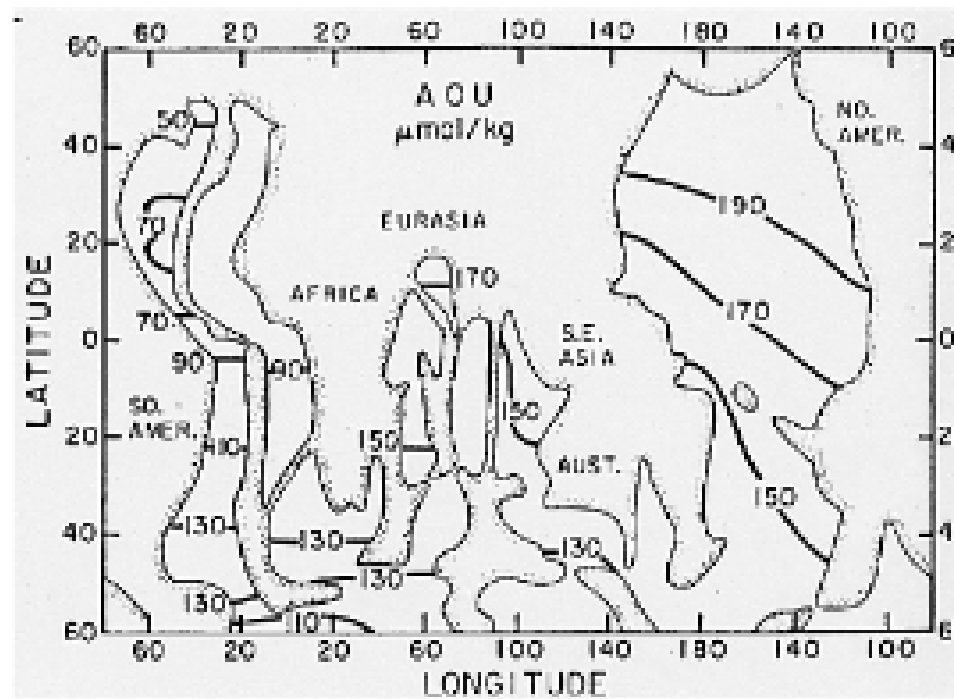
$$AOU = O_2^s - O_2$$

where:  $O_2^s$  = value of  $O_2$  the water would have if it was in equilibrium with the atmosphere at the temperature and salinity of the water. This is called saturation. This implies that all waters are in equilibrium with the atmosphere (100% saturated) when they sink to become the deep ocean water.  $O_2$  is the dissolved oxygen actually measured in the same water sample.

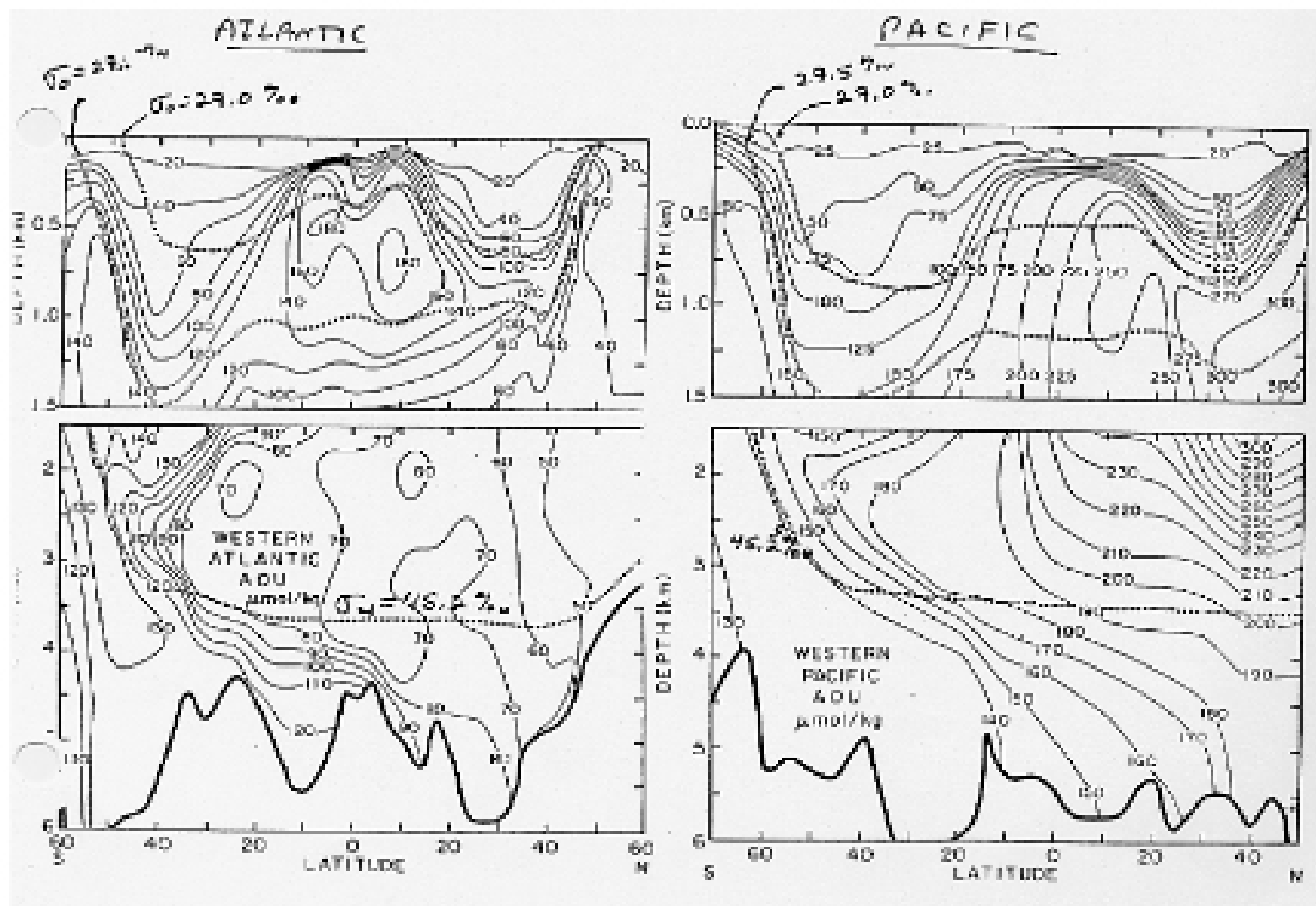
The distribution of AOU throughout the ocean at 4000m is shown in the attached **Figure 9-1** (from Broecker and Peng). The lowest values ( $50 \mu\text{mol kg}^{-1}$ ) are in the North Atlantic. The highest values ( $>190 \mu\text{mol kg}^{-1}$ ) are in the oldest water in the North Pacific.

North-south sections of AOU in the western Atlantic and western Pacific Oceans are shown in **Figures 9-2a,b**. Some key intermediate and deep density surfaces ( $\sigma_\theta = 27.0$ ,  $27.5$  and  $\sigma_4 = 45.2$ ) are indicated.

**Fig 9-1** Distribution of AOU at 4000m in the world's oceans. Lowest values are in the Atlantic and the largest ( $>190 \mu\text{mol kg}^{-1}$ ) are in the Pacific. (from Broecker and Peng, 1982).



**Fig 9-2 a,b** Sections of AOU through the Atlantic and Pacific



### Regenerated Nutrients

Once you've calculated the AOU in a water sample you can calculate the  $\text{CO}_2$ ,  $\text{HNO}_3$  and  $\text{H}_3\text{PO}_4$  released by respiration.

$$\begin{aligned} 1 \text{ mol O}_2 \text{ consumed} &= 106/138 \text{ mol CO}_2 + 16/138 \text{ mol HNO}_3 + 1/138 \text{ mol H}_3\text{PO}_4 \\ &= 0.77 \text{ CO}_2 + 0.12 \text{ HNO}_3 + 0.0072 \text{ H}_3\text{PO}_4 \end{aligned}$$

### Preformed Nutrients

Preformed nutrients are those initially present in seawater at the time of downwelling. Hence, preformed nutrient = total nutrient - regenerated nutrient.

An example from Park (1967) of the relationships between AOU: $\text{PO}_4$ : $\text{NO}_3$ :Si is shown in the Fig. 9-3. The location of these profiles is at  $44^\circ\text{N}$ ;  $127^\circ\text{W}$  (off the coast of Oregon). The nutrients of oxidative origin ( $P_{\text{ox}}$ ) were calculated from AOU and subtracted from total nutrients to get the preformed nutrient concentrations.

Preformed nutrients are characteristic of waters originating in different regions and hence can be used as water mass tracers (e.g. Broecker, 1974; Broecker et al, 1985). Broecker (1974) originally proposed that  $\text{O}_2$  and  $\text{NO}_3$  and  $\text{O}_2$  and  $\text{PO}_4$  data can be combined in such a way that the alteration by respiration is cancelled. From the original RKR respiration equation given above, roughly 1/9 mole of N is released as  $\text{NO}_3$  and 1/135 mole of P is released as  $\text{PO}_4$  for each mole of  $\text{O}_2$  consumed. Thus the sum of  $9\text{NO}_3 + \text{O}_2$  was defined as "NO" and  $135\text{PO}_4 + \text{O}_2$  as "PO". Both NO and PO should be nearly conservative tracers. An example of  $\text{NO}_3$ ,  $\text{O}_2$  and "NO" versus salinity for a single station in the western basin of the South Atlantic is shown in Fig. 9-4. You can see that  $\text{O}_2$  and  $\text{NO}_3$  show curvature with salinity but "NO" varies linearly, or conservatively, between two end members that represent the core of the AAIW (at 1150m) and the top of the NADW (at 2270m).

### Revised Stoichiometric Ratios

Isopycnal surfaces are surfaces of constant density. Potential temperature and salinity are perfectly correlated on isopycnal surfaces. In the ocean most mixing and transport occur on isopycnal surfaces. The winter surface outcrops of various isopycnal surfaces in the North Atlantic Ocean are shown in Fig. 9-5. The densest waters that outcrop at the surface have a density of  $\sigma_\theta = 27.6$ . The north-south distribution of density surfaces in the 0 - 1500m depth range of the western side of the Atlantic, Indian and Pacific Oceans are shown in Fig. 9-6. The density surfaces shoal and outcrop at high latitude, then sink to maximum depths at mid-latitudes. They tend to shoal again near the equator where there is upwelling. To a first approximation the distributions of tritium ( $^3\text{H}$ ), which is a tracer for water, which has an atmospheric source, follow these density surfaces.

Takahashi et al (1985) first argued that the correct approach for determining stoichiometric regeneration ratios was to utilize data along isopycnal surfaces. The change in  $\text{NO}_3$  and  $\text{PO}_4$  concentrations along isopycnal surfaces away from the surface outcrop can be estimated from: