Prospects

What is the likelihood of discovering more vibrations? The possibility of more complex motion always exists, and other vibrations can be calculated in RPA. However, all the other modes of particle displacement have a higher mean frequency. The damping is predicted to be much larger, both within RPA and due to the coupling to other degrees of freedom. Thus it is unlikely that a coherent mode, containing most of the sum rule, will be resolved as a distinct peak. In the inelastic scattering experiments, the spectrum contains a smooth background together with the presently known peaks, as may be seen in Figs 3 and 4. Part of this background at least must be due to highly damped vibrations of greater complexity. Without some specific measurable characteristic, however, it is not possible to identify them as such.

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articles

A 30,000-yr isotope climatic record from Antarctic ice

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Simple glaciological conditions at Dome C in east Antarctica have made possible a more detailed and accurate interpretation of an ice core to 950 m depth spanning some 32,000 yr than that obtained from earlier ice cores. Dated events in comparable marine core has enabled the reduction of accumulation rate during the last ice age to be estimated. Climatic events recorded in the ice core indicate that the warmest Holocene period in the Southern Hemisphere occurred at an earlier date than in the Northern Hemisphere.

ALTHOUGH the stable isotopic composition ($^{18}O/^{16}O$, D/H) of polar ice sheets can provide a continuous record of past climatic conditions, the interpretation of the isotope profiles in terms of climatic temperature changes over a time scale ranging back to the last ice age is complicated by several factors including: (1) the determination of the time scale; (2) the influence of the elevation at which the snow was deposited and of ice sheet stability; (3) the establishment of a transfer function from isotopic δ ratio to temperature in past conditions. Despite these limitations the study of ice cores has already provided very useful palaeoclimatic data[1–5]. Similar limitations occur in the interpretation of other indirect sources of climatic data.

As isotopic records from areas near an ice divide are simpler to interpret, thermal core drilling to 906 m depth was carried out at Dome C (74° 39′ S; 124° 10′ E; elevation: 3,240 m, mean annual temperature: −53.5 °C, Fig. 1) during the 1977–78 Antarctic field season as part of the International Antarctic Glaciological project[6].

The isotope profile

The core recovery was about 98%. Sampling was carried out in the field by cutting a continuous slice from along the length of cleaned ice core. The $^{18}O/^{16}O$ ratios were measured with a fully automatic double mass spectrometer with an accuracy of 0.15‰ in the δ scale relative to the V. SMOW. The results averaged for samples of about 4 m length are plotted in Fig. 2 with depths expressed in metres of ice equivalent to take account of the

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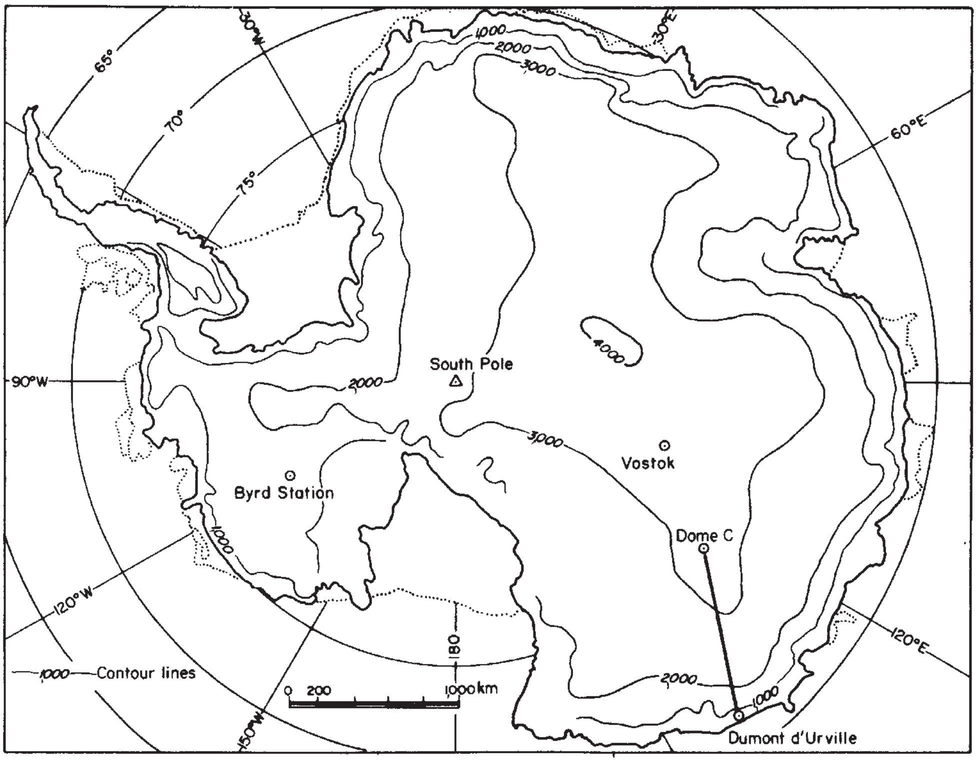

Fig. 1 Map of Antarctica showing the location of several stations (including the drilling sites) and of the traverse from Dumont d'Urville towards Dome C (dashed line). Recent data[32] indicate a dome-like shape around the Dome C site.

lower densities in the upper firn layers. Each increment samples a time interval a little longer than one century. A smooth curve was drawn from adjustment on the basis of a spline function of the order of 2 with a parameter of adjustment $\rho = 0.001$. The main features of the δ curve indicate relatively uniform conditions down to 381 m, with a mean value of −49.9‰ (standard deviation $\sigma = 0.46$) followed by a rapid decrease down to 510 m. From then and down to 687 m δ fluctuates around −55.3% ($\sigma = 0.37$); isotopic fluctuations and presumably those of climate are of similar magnitude in this depth interval to those of the upper 381 m, when mean values over the order of 100 yr are considered. The general trend is then towards less negative values down to the bottom (875 m, ice equivalent). Significant events are superimposed on these main features; some of these are indicated in Fig. 2.

Ice core chronology

Different methods may be used for dating deep ice cores[7,8] including measurements of radioactive decay, periodic changes of various parameters, and specific horizons dated from other sources. Some of these studies will be performed later, although they may not be successful in the low accumulation of the Dome C area. At present, a preliminary time scale can be calculated from the simplest ice flow model[9] which assumes a uniform vertical strain rate and a constant value of snow accumulation. Although this model has other limitations[10] it can be applied with reasonable confidence to ice dome areas when bedrock is flat and when considering the upper part of the ice sheet; both these conditions are fulfilled for the Dome C ice core. However, a large uncertainty results from the assumption of steady-state conditions over a time range which goes back to the last ice age. This is particularly true for the rate of snow accumulation which has so far been determined only over the past 25 yr from the radioactive layers formed from nuclear tests (0.037 m ice $yr^{-1} \pm 0.004$). The age of the ice (t yr) with depth can be calculated from

$$t = \frac{H}{\lambda} \ln \frac{H}{H - z} \qquad (1)$$

where λ is the accumulation rate, H the total ice thickness (3,400 m) and z the depth, all expressed in metres of ice equivalent. The ages so calculated are given in Fig. 2 (numbers in parentheses).

If the main isotopic features reflect climatic changes, we can try to get chronological information from climatic records obtained from other sources. Data from continental areas of the Southern Hemisphere are scarce but some reference data may be obtained from marine sediment cores which provide a climatic record of surface seawater temperature, assuming that typical climatic events are synchronous with those recorded in Antarctic snows. Previous results[11] from core MD 73025 (Indian Ocean; 43° 49′ S, 51° 19′ E) have been provided by J. C. Duplessy and G. Delibrias (personal communication). Eight ^{14}C ages were determined on the top 3.5 m of this core and the comparison of pelagic and benthic foraminifera isotopic profiles shows that some of the characteristic events recorded in the Dome C ice core are also apparent in this marine core. The selected climatic features can be seen in Fig. 2; depth in both cores with the three ^{14}C interpolated marine sediments ages are given in Table 1. According to Duplessy these climatic features can be recognised in several northern and southern ocean deep-sea cores having a high accumulation rate with roughly the same ^{14}C ages within experimental errors.

Table 1 Depth and interpolated marine sediment age of Dome C and MO 73025 cores

Dome C ice core Depth (m ice equivalent)	MD 73025 marine core Depth (m)	^{14}C (yr)
381	2.36 ± 0.02	10,550 ± 300
437	2.82 ± 0.02	12,600 ± 300
510	3.10 ± 0.05	15,500 ± 500

Snow accumulation rate changes

From these marine dates, we calculate by equation (1) a mean rate of snow accumulation in different ^{14}C time intervals. This gives 0.038 m yr^{-1} for the past 10,550 yr, 0.031 m yr^{-1} from 10,550 to 12,600 yr BP and 0.029 m yr^{-1} from 12,600 to 15,500 yr BP.

Although there are many uncertainties in such an approach and the accuracy of the accumulation figures may not be better than 10–15% (Table 1), the two last values are based on differences between ^{14}C ages, thus reducing problems of a possible time lag between marine and ice records and of errors of ^{14}C dating.

The rate of accumulation over the past 10,000 yr agrees very well with the value (0.037) measured over the past 25 yr. Since we know there are short-term changes in the accumulation rate this may seem to be a coincidence. However, the conclusion that present-day accumulation rates are typical of those for the past few thousand years has already been stated[12] for both the Greenland and Antarctic ice sheets from dating[7] and inferred from observed temperature–depth gradients[13]. Our figures suggest a decrease of the rate of snow accumulation associated with colder climatic conditions, the accumulation 14,000 yr ago being about 75% of the present value.

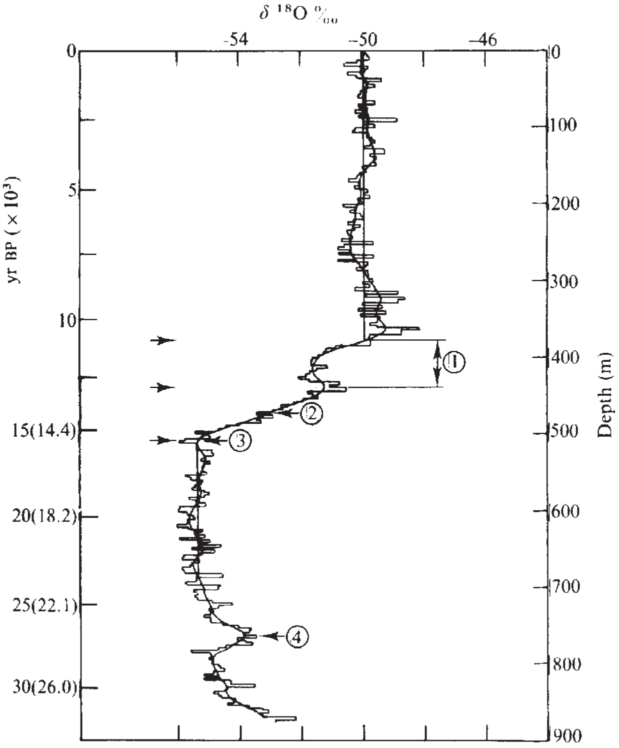

Fig. 2 δ record from the Dome C ice core plotted for about 4-m long increments and smoothed curve. The straight lines indicate mean values over the past 11,000 yr and during the late part of the last ice age. Significant isotopic events are numbered on the right hand side. On the left hand side arrows indicate reference levels corresponding to ^{14}C dates obtained from marine sediments. All depths are expressed in metres of ice equivalent. Ages are estimated from a simple ice flow model assuming a variable or constant (numbers in brackets) rate of accumulation.

Note that the mean annual accumulation in Antarctica is primarily governed by the amount of water vapour that can be carried in the atmosphere circulating above the surface inversion layer[12]. This is governed by the atmospheric temperature and by the saturation water vapour pressure over ice above the inversion. For condensation temperatures of around −40 °C in the Dome C area, a ratio of 0.75 for the vapour pressures corresponds to a condensation temperature change of about 3 °C.

The study of precipitation[14] at the South Pole gives a gradient of $d\delta^{18}O/dT_c = 1.4‰$ per °C, where T_c is the condensation temperature obtained from radiosonde measurements of cloud temperature. Analysis of snow samples collected between Dumont d'Urville and Dome C (Fig. 1) has provided an empirical relationship[15] between the mean surface temperature T_m and the isotopic deuterium content of $\delta D‰ = 6.04\ T_m$ (°C) -51 equivalent to $d\delta^{18}O/dT_m = 0.75‰$ per °C. In Antarctica mean temperatures above the atmospheric inversion layer (T_i) are similar to the condensation temperatures[12], while the inversion strength is linearly related[16] to mean surface temperature. A survey of existing data for eight stations with T_m values in the range from −10 to −56 °C gives $dT_i/dT_m = 0.67$ with a correlation coefficient of 0.99. Assuming $T_i = T_c$ leads us to interpret our traverse results of equivalent to $d\delta^{18}O/dT_c = 1.1‰$ per °C, a value which is in reasonable agreement with that obtained at the South Pole.

A change of 3 °C for the temperature of condensation previously deduced from accumulation data would lead to a change of 3.3‰ between ice recently deposited and that deposited 14,000 yr ago. The agreement with the value obtained from the Dome C record (3.1‰) is very satisfactory considering the large number of assumptions involved, which confirms conclusions about snow accumulation changes inferred from the Indian Ocean marine core data.

To obtain a continuous time scale we use relation (1) with a constant rate of accumulation (0.037 m ice yr^{-1}) down to a depth of 381 m, and then decreasing to 75% of this value down to 510 m; this last figure is used down to 687 m. Further down it does not make a significant difference whether or not we assume an increasing accumulation linked to the $\delta^{18}O‰$ change. The calculated time scale is given in Fig. 2. According to this model the age of the bottom of the 905 m deep ice core is about 32,000 yr.

Climatic interpretation

Two other ice cores with depths comparable or greater than that from Dome C have been recovered in Antarctica (Fig. 1). At Byrd Station (80° 01′ S, 119° 31′ W) the 2,160-m deep ice core covers a much longer time period and the isotopic change towards much more negative values is located between 1,000 and 1,400 m. The location and characteristics of the Vostok ice core are more comparable to those of Dome C one, the isotopic change at Vostok occurring between 250 and 375 m. Part of the published[2,17] isotope profiles plotted against ages are shown in Fig. 3.

Comparison of Figs 2 and 3 shows similar trends in the three δ profiles. Event 1 from Fig. 2 is well marked in all three locations and there is a parallelism between the Byrd and Dome C isotopic profiles down to event 3. Events 1, 2 and 3 are also apparent in the Vostok ice core when looking at a more detailed isotope profile[4], although there is a large scattering of the δ values probably due to discontinuous sampling technique. The less negative values of event 4 are also observed in the other ice cores.

Dating problems make precise age comparisons difficult, but it must be pointed out that the depths of the four events in the Vostok and Byrd ice cores relative to that of the Dome C record show a rather constant value. For the Vostok ice core where $\lambda = 0.024$ m ice yr^{-1} this value is similar to the ratio of the present rates of snow accumulation; this is not the case for the Byrd ice core, probably due to the very complicated regime in this area.

It appears then that the main characteristics of the δ profiles are similar for east Antarctica (Dome C, Vostok) and west Antarctica. Thus they probably reflect climatic events which affected both east and west Antarctica, rather than any ice sheet instability[18] because these two distinct areas have very different glaciological regimes. Furthermore, isotopic events observed in the Dome C ice core are also seen in the record of sea-surface temperature changes in marine sediments.

The ice core chronology from Dome C indicates a relatively cold period from 11,000 to 13,000 yr BP; this is shown also in the record of past temperature of the New Guinea Highland summarised in ref. 19 from careful pollen analysis. The Dome C fluctuations observed during this period could be equivalent to the European Dryas–Bölling–Alleroïd stages depicted in the Camp Century (Greenland) ice core[1] although dating uncertainties prevent any conclusion about their synchronism.

The last major glacial peak occurs around 16,000 yr BP, which agrees with the much less detailed pollen record from New Guinea[19] and the indication from various sources[20–22] that continental warming had probably begun in the Australian–New Guinea area by $16{,}000 \pm 2{,}000$ BP. This worldwide climatic feature has been found in many marine sediment cores both in the Northern and Southern Hemispheres[23,24] with an age of about 18,000 yr which may vary locally by several thousand years.

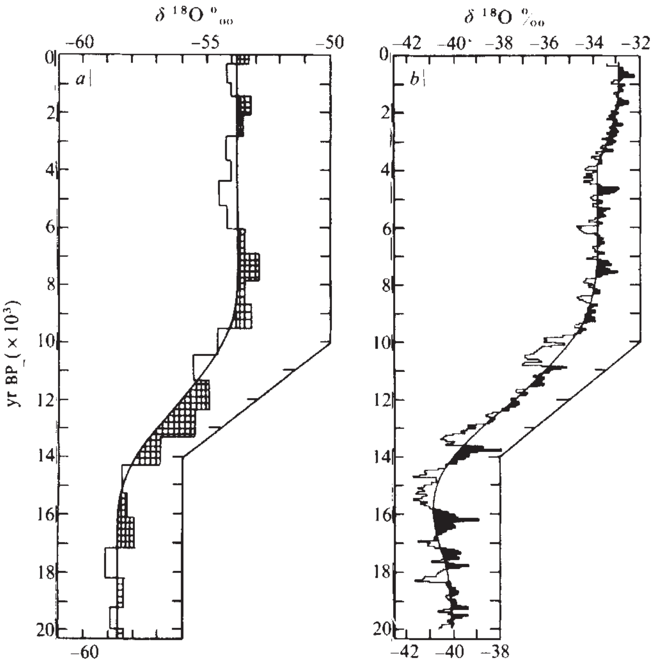

Fig. 3 The Vostok (*a*) and Byrd (*b*) δ records during the past 20,000 yr from ref. 2 (Byrd) and ref. 17 (Vostok). The δ values exceeding the smoothed curves are shown in black.

The warmer conditions shown on the Dome C δ record before 25,000 yr BP also agrees with conditions observed around this time in New Guinea by pollen analysis[22] and with North America glacial chronology[25].

The isotopic shift associated with the transition from the coldest part of the ice age towards mean climatic conditions from the last 10,000 yr is 5‰ for Vostok[4], 7‰ for Byrd[2] and 5.4‰ for Dome C. When values for Vostok and Byrd station are corrected[12] to allow for the origin further inland of the deeper ice and for a higher surface elevation of the ice sheet during the last ice age, the δ^{18}O‰ change associated with climatic changes would become 4.5 (Vostok) and 5.3 (Byrd). Although corrections for the same factors at Dome C could be made if measurements of total gas content[26] were available, severe cracking in the ice core from Dome C has made such a study impracticable. Since Dome C is at present a centre of outflow and if we assume that the form of the ice sheet has changed little over the past 10,000 yr, then the δ values in the upper 381 m of ice core should provide a useful climatic record over this period.

However, at the end of the last ice age, a lower sea level may have resulted in a slightly expanded ice sheet over east Antarctica, leading to greater ice thicknesses and increased surface elevations. These effects would be much greater in coastal areas, while in central east Antarctica estimates of the consequent surface rise are around 100 m (refs 4, 28). Such a rise, together with the present δ-elevation gradient observed in the Dome C area[15] would involve a correction of 1.5‰ to allow for changed elevation during the ice age. Thus it seems that the climatic change at the end of the last ice age was similar at Dome C, Vostok and Byrd, but further studies of ice thickness changes at Dome C are needed to confirm this conclusion.

The isotopic shift has still to be corrected for modifications of oceanic parameters during the last ice age. The most obvious is the change of mean isotopic value of oceanic waters. Due to the storage of water in δ-depleted ice sheets an increase of 1.6 δ^{18}O‰ with reference to the present conditions has been estimated[11]. This effect will probably be reflected in antarctic precipitation but from previous discussion it could be compensated by greater ice thickness during the last ice age.

The isotopic composition of ice cores, although principally governed by the temperature of condensation, can be affected by many other parameters[1]. However, it has been suggested[12] that the present empirical relationship between mean surface temperature and isotopic content may not have been too different in Antarctica during the last ice age. As previously mentioned a 0.75‰ per °C δ^{18}O change was obtained in this area for a large range of temperatures, which leads to a tentative estimation of a difference of about 7 °C for the surface temperature between the coldest part of the ice age and the present climate. From previous discussions the suggested changes for the temperatures of condensation and above the inversion layer are reduced by about half.

Although there are many assumptions in such interpretations, they agree with the results of atmospheric modelling of the ice age (18,000 yr BP) climate[29,30] which also indicate lower precipitation during cold conditions comparable with those described here. For the past 11,000 yr the climatic record inferred from the δ profiles (Fig. 2) shows that the warmest conditions occurred during the period 11,000–8,000 yr BP and the coldest conditions from about 8,000 to 4,000 yr ago. For the past 4,000 yr we observe a moderately warm period followed by a slight trend towards less positive δ values, with present surface conditions close to the mean value for Holocene.

Figure 3 shows that the Vostok δ profile indicates the same succession of climatic periods with a slightly different chronology. However, the Byrd record shows no apparent agreement with this sequence[2] due to the difficulty of interpreting its complicated glaciological regime.

The lack of comparative data from the Southern Hemisphere makes it difficult to estimate the spatial representativity of the Dome C Holocene record. During this period, the climatic evolution inferred from our δ profile gives a different picture from that obtained from New Guinea pollen analysis[19,20]. The warmest period (11,000–8,000 yr BP) that we observed at the end of the transition from the last ice age is much earlier than the usual estimates for the so-called hypsithermal in the Northern Hemisphere. This is, however, in excellent agreement with the fact that the warmest sea surface temperature since the last glacial age[31] in the Southern Hemisphere have a radiocarbon date of 9,400 yr (24).

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Shelf sea fronts' adjustments revealed by satellite IR imagery

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IR imagery from NOAA 5 has been used to study the movement of shelf sea fronts. After the elimination of tidal advection, the fronts show remarkable consistency of position and do not adjust to the strong semi-monthly variation in tidal stirring which indicates the operation of a feedback process in vertical mixing.

IR IMAGERY from NOAA 4/5 has provided convincing evidence of the role of tidal mixing in controlling the distribution of sea surface temperatures in the European shelf seas. In particular it has been shown that the positions of the surface manifestations of the shelf sea fronts are closely parallel to contours of the h/u^3 parameter[1,2], as previously predicted[3].

As the archive of imagery grows, it should be possible to use it to answer more detailed questions about the evolution of the fronts and the processes occurring on them. Indications have already been obtained from several images of the development of large scale (~25 km) instabilities in the along front flow which are clearly important as agencies of cross-frontal mixing[1,4].

In this article we attempt a more systematic analysis of two years data from NOAA 5 to investigate the extent to which the mean positions of the fronts adjust to variations in both the rate of tidal mixing and the surface heat flux.

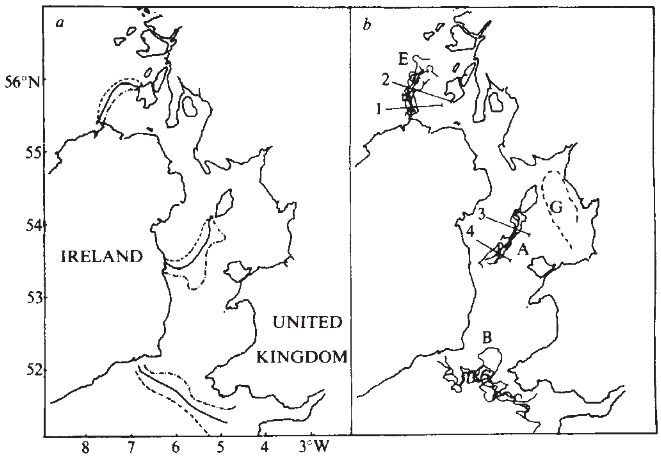

Fig. 1 *a*, Contours of $\log_{10} h/u_s^3$ –·–· 1.5; —— 1.9; - - - 2.2. h Is the depth (m) and u_s is the tidal stream amplitude at springs (m s^{-1}). The separation of the contours 1.5 and 2.2 indicates the range of hypothetical adjustment between neaps and springs. Contour shape based on Pingree and Griffiths[2]. *b*, Composite of all frontal positions observed in July 1977 from NOAA 5 imagery. Fronts are labelled as follows: A, Western Irish Sea (WIS); B, Celtic Sea; E, Islay; G, Liverpool Bay. Reference lines, along which displacements were measured, are numbered.

Equilibrium adjustment

We particularly wish to examine the possibility that the mean frontal positions vary during the semi-monthly range cycle of the tides. As the amplitude of the tidal streams increases by a factor ~1.8 from neaps to springs, the intensity of tidal dissipation, which is proportional to u^3, will rise by a factor ~5.8. The balance between surface heating and tidal and wind stirring at a front may be expressed in a constant efficiency model[1] as

$$R = 1 - a\left(\frac{\overline{u^3}}{Qh}\right) - b\left(\frac{\overline{w^3}}{Qh}\right) = 0$$

where h is the depth; u the tidal stream velocity; w the windspeed; Q the surface heating rate, and the parameters a and b, which contain the efficiencies of tide and wind mixing, may be assumed constant. The overbar denotes a time average.

If we further assume that Q and $\overline{w^3}$ are spatially uniform over a particular frontal region, the 'equilibrium positions' should be defined by

$$\chi = \frac{h}{\overline{u^3}} = \text{constant}$$

and, for slow changes in $\overline{u^3}$, the front would move to the appropriate contour of χ. This hypothetical adjustment, for a range change corresponding to the phase inequality cycle is shown in Fig. 1*a*. In the case of the Islay front, for example, the equilibrium displacement would be ~20 km. In other frontal

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