I.O.S.

BENTHIC BOUNDARY LAYER

IOS OBSERVATIONAL AND MODELLING PROGRAMME

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BY

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OCEAN DISPOSAL OF HIGH LEVEL RADIOACTIVE WASTE
A RESEARCH REPORT PREPARED FOR THE DEPARTMENT
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NATURAL ENVIRONMENT RESEARCH COUNCIL

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Abstract (100-200 words as desired)

Near bottom currents, measured at three sites in the N.E. Atlantic, reveal the eddying characteristics of the flow. Eddies develop, migrate and decay in ways best revealed by numerical modelling simulations. Eddies control the thickness of the bottom mixed layer by (frictionally) accumulating and thickening or spreading and thinning the bottom waters. At the boundaries of eddies benthic fronts form providing a path for upward displacement of the bottom water.

An experiment designed to estimate vertical (diapycnal) diffusivity is performed. The flux of heat into the bottom of the Iberian basin through Discovery Gap is deduced from year long current measurements. The flux is supposed balanced by geothermal heating through the sea floor and diapycnal diffusion in the water. A diffusivity of 1.5-4 $cm^2\ s^{-1}$ is derived for the bottom few hundred meters of the deep ocean.

Experiments to etimate horizontal (isopycnal) diffusivity are described. Diffusion WITHIN oceanic eddies is shown to obey a law found for atmospheric cyclones and anticyclones. Diffusion BY eddies is determined largely by their energy and in the abyssal N.E. Atlantic a value of $10^6\ cm^2 s^{-1}$ is determined.

If a tracer is discharged from the sea bed the volume of sea water in which it is found increases with time and after 20 years will fill an ocean basin of side 1000 km to a depth of only 1-2 km.

Keywords:
299, 93, 94, 126, 155 DoE sponsored research, Disposal on/under deep ocean bed, Ocean circulation/dispersal, Aquatic dispersion.

This work has been commissioned by the Department of the Environment as part of its radioactive waste management research programme. The results will be used in the formulation of Government policy, but at this stage they do not necessarily represent Government policy.

Contents

PREFACE

The research described in this report is concerned with the scientific assessment of the feasibility of the disposal of heat generating radioactive waste (HGW) into the deep sea environment. It deals with the deep ocean water column and is aimed at understanding the initial mechanisms of dispersion of radionuclides introduced into the benthic boundary layer (BBL). This layer, adjacent to the seabed and varying from 10-100m in thickness is caused by friction between the moving ocean water and the stationary ocean bottom. Within the BBL turbulent mixing is sufficiently strong that the properties such as density (and by analogy a radionuclide source term), are rendered uniform in the vertical. Above it the density decreases with height and vertical exchange is suppressed.

The Natural Environment Research Council, through the Institute of Oceanographic Sciences, had a contract with the Department of the Environment (DoE, DGR481/176) to examine processes within the BBL both by direct measurement within the deep ocean and also by numerical modelling. The emphasis of the investigations has been placed on studying those processes relevant to radionuclide dispersal within the BBL in order that realistic predictive models can be developed.

Interim reports have already been published on the observational (DoE/RW/83.067) and modelling (DoE/RW/83.070) aspects of the research: this report combines both aspects and is the final one under the contract.

INTRODUCTION - RESOURCES AND METHODS EMPLOYED

(a) Observational Programme

Measurements within and above the benthic boundary layer have been made at three sites in the N. E. Atlantic. Their locations are shown on Figure 1 and all had water depths of between 4500 and 5300m. The first site was on the Madeira abyssal plain in a generally flat area with isolated small abyssal hills about 400m high. The third site was 150 miles west of Madeira on a gently sloping part of the lower continental rise. The second site was in Discovery Gap, a narrow and rugged channel in the East Azores Fracture Zone which communicates between the deepest levels of the Madeira and Iberian Abyssal Plains. The varying topographic nature of these sites ensures that the benthic boundary layer and its dispersion characteristics differ widely.

Flow was measured using moored self-recording current meters in groups of about 20 at heights between 1m and 1000m above the sea bed. The duration of the measurements was between 4 months and 1 year. Currents were also measured by acoustic tracking of a cluster of between 5 and 10 neutrally buoyant floats: such floats follow the water at a fixed depth and were positioned 100-300 metres above the bottom. Tracking of the floats requires the presence of a ship nearby and hence was limited to 10 to 20 days at each site.

Observations of water temperature are invaluable in revealing the vertical extent of the boundary layer and the nature of processes within it. Temperature measurements were made with moored current meters (which also determined the flow) and with platinum resistance thermometers lowered from the ship.

The extent and scope of the experimental programme is summarised in the accompanying table. The resources brought to bear on each experiment were considerable and their cost has been shared by NERC and DoE.

Table 1

Details of the BBL experimental programme

Site 1 Madeira Abyssal Plain

Location 33^0N 22^0W: water depth 5300m approximately

No. of moorings deployed 7. No. of current meters 21. Duration 7 months.

No. of neutrally buoyant floats launched 12.

No. of lowered profiles within radius of 50km 20.

RRS Discovery Cruises 114, 117, 122.

Site 2 Discovery Gap

Location 37^0N 16^0W: water depth 4700m approximately.

No. of moorings deployed 8. No. of current meters 12. Duration 11 months.

No. of neutrally buoyant floats launched 10.

No. of lowered profiles within region 25.

RRS Discovery Cruises 130, 138.

Site 3 Madeira Continental Rise

Location 33^0N 20^oW: water depth 4500m approximately.

No. of moorings deployed 7. No. of current meters 20. Duration 4 months.

No. of neutrally buoyant floats launched 10.

No. of lowered profiles within 50km radius 20.

RRS Discovery Cruises 138, 139, 141.

(b) Numerical Modelling

The aim of the modelling programme has been to study some of the processes occurring close to the seabed and to provide a theoretical framework within which to interpret and extend the results obtained from the observational programme. Two approaches have been taken. The first investigates the development of the mixed layer under a horizontally uniform current. A sophisticated turbulence model has been used. The model provides estimates of the expected height of the layer and how long it will take to reach this height together with the vertical variations of the flow and turbulence quantities.

This proved a stepping-stone to a second model in which the flow varies both in space and time. This allows the study of flows characteristic of the deep ocean and the horizontal dispersion of a tracer by such flows. The second model required the use of the very fast and powerful CRAY computer.

THE STRUCTURE OF THE FLOW IN THE DEEP OCEAN

Two of the experiments conducted under this contract permitted currents to be mapped in the deep ocean, see Figures 2 and 3. Both figures show streamlines of the flow, or isobars, drawn at a height of 10m above the seabed at 10 day intervals. Although such maps have been drawn for deep levels in the N.W. Atlantic[1], these are the first drawn for abyssal levels in the Eastern atlantic.

The maps reveal a series of eddies, the precise oceanic counterpart of atmospheric cyclones, which intensify or weaken and also migrate. Between atmosphere and ocean there is a striking contrast in both size and strength of the eddy circulation; in the atmosphere eddies have a diameter of 2000km with winds of 20m/sec, in the deep ocean their diameter is only 50km with currents of about 10cm/sec. The development, migration and decay of these eddies is also well described by numerical simulations[28].

Currents measured on a mooring largely reflect the strength and passage of these eddies: measurements of their energy at the three experimental sites are presented in Table 2. It will be noted that the quietest location is on the abyssal plain[2] with the most energetic region on the lower rise: in Discovery Gap the flow is intermediate in energy. Only at the latter location is there a significant long term mean current[3].

Table 2

Energy of the eddy flow in the N.E. Atlantic

Site and bottom depth	Duration, days	Height above bottom, m	Number of Instruments	Kinetic Energy cm^2s^{-2}
Madeira Abyssal Plain 5300m	224*	10	7	3.6
		100	7	2.7
		600	3	1.6
Madeira Rise 4650m	120	10	4 (1)	27.5 (16)
		100	4 (1)	25 (14)
		600	1	10
		1500	1	4.5
Discovery Gap 5000m	325+	10	2	6.1
		300	3	4.9
		600	2	3.9

*+ More complete description of the data sources is found elsewhere.[2,3]

A comprehensive survey of near bottom currents between 20 and 50° N in the N.E. Atlantic[4] reveals that over large areas of abyssal plain the eddy kinetic energy is close to 1-2 $cm^2 s^{-2}$, similar to that reported here. The data of Table 2 also indicates that the eddy circulation is most intense near the top of the BBL with energy falling to one-half of the peak value at heights between 500 and 1000m above the seabed.

The most abrupt height variations in current occur within the BBL itself, within the lowest meter above the seabed, see Figure 4. On the bottom the current vanishes altogether and at 1m height, as the numerical modelling shows, its strength is about 80% of that at 10m[5]. At some height between 10 and 100m the direct frictional influence of the seabed can no longer be detected. The current is not quite unidirectional within the BBL but because of the increasing importance of friction turns towards areas of low pressure, through an angle of 5-15 degrees[5,6]. It is also, of course, turbulent.

A time series of current measurements, see Figure 5, not only reflects the development and movement of ocean eddies. In addition to slow time variations with periods ranging from 5 to 100 days, see Figure 6, there are also present relatively rapid variations with well defined periods. These are tides and inertial oscillations, both of which influence the structure of the BBL.

The principal tidal current, the lunar semi-diurnal component, has a period of 12.42 hours and is found to be approximately constant over the whole water column: it is closely associated with changes in sea level. At each of the three sites the tidal currents are approximately in phase, but differences can be seen amongst the sites (Table 3). The tidal amplitude is between 3 and 4 cm/s in the region studied, on the abyssal plain quite comparable with the eddy currents: the orientation of the tidal ellipse is about 045°. These latter properties are quite well predicted by the tidal model of Schwiderski[7] which gives confidence in its predictive powers elsewhere.

Inertial oscillations have a frequency equal to twice the vertical component of the angular spin of the earth and an amplitude of about 1 cm/s, see Table 4. Because this is considerably smaller than the tidal component their effects on the boundary layer are correspondingly weaker.

Table 3

Lunar semi-diurnal tidal currents at exptl. sites in the N.E. Atlantic

Experimental measurement / predictions according to Schwiderski[7]

Site	Location	ellipse axis, cms^{-1} major	ellipse axis, cms^{-1} minor	orientation, °T	phase, deg
Abyssal Plain	33N 22W	3.4	−0.6	44	−2
		2.7	+0.2	37	13
Lower Rise	32N 20W	3.2	+0.4	39	−40
		3.9	−0.4	44	3
Discovery Gap	37N 15W	3.9	−0.2	46	14
		3.0	+0.3	34	35

Table 4

Amplitude of inertial oscillations

Site	Location	Period, hrs	Height above bottom, m	Number of Instruments	Amplitude, cms^{-1}
Abyssal Plain	33N 22W	19.5 - 23.0	10	7	1.3
			100	7	1.4
			600	3	0.9
Lower Rise	32N 20W	17.7 - 24.0	10	4	0.9
			100	4	1.2
Discovery Gap	37N 15W	18.0 - 20.0	10	2	0.45
			300	3	0.5
			600	2	0.55

VERTICAL OR DIAPYCNAL MIXING IN THE OCEAN

Because the ocean is a very thin shell of fluid on the earth's surface its motions and processes in (locally) vertical and horizontal directions have quite different properties. The former are considered in this section and the latter in the subsequent one.

(a) Thickness of the bottom mixed layer

Temperature profiles obtained from a lowered recording thermometer reveal a layer of uniform potential temperature immediately above the seabed[8]. The layer is commonly known as the bottom mixed layer and its thickness is quite variable. On the Madeira abyssal plain the thickness ranged between 20 and 120m, with a mean of 55m from 20 lowerings. On the Madeira rise it ranged between 7 and 145m with a mean of 45m from 28 lowerings.

Models of the benthic boundary layer which consider the development of the layer and its vertical structure under a horizontally uniform current cannot account for either the mean thickness or the variation of layer height[5]. Observations[8] show that the height of the layer is not correlated with the strength of the flow above the BBL as is implied by one dimensional models. A numerical model which examines the interaction of the benthic boundary layer with the deep ocean eddies described in the previous section has been developed at I.O.S.[9]. One of the main results of the model is the control by eddies of the height of the benthic boundary layer. Due to convergences and divergences produced by the eddies the height of the layer is distorted, see Figure 7. The variations in the height of the layer predicted by the model are very similar to those observed. The height of the benthic boundary layer is such that it is not expected that bottom generated turbulence will penetrate the full extent of the layer. At present we suppose that the upper parts of a deep mixed layer of height 100m or more are quiescent or 'fossil' where active mixing is not taking place.

(b) Benthic Fronts

A time series of temperature measurements made in the benthic boundary layer reveals the occurrence of occasional jumps in temperature[10]. Although the magnitude of these jumps is only .002 to .005°C they have the character of fronts. Vertical cross-sections (Figure 8) reveal that the frontal surfaces are about 300m in width, tilt at a mean angle of 10 degrees from the horizontal and meander over horizontal distances of 1-5km. The phenomena are termed benthic fronts. One such front on the Maderia abyssal plain extended at least 30km in the horizontal and 600m in the vertical. The frequency of frontal passage appears to be several per year. Such fronts are capable of sweeping up water from the benthic boundary layer on one side and ejecting it at heights of 50 to 200m above bottom on the other side. Although we have not observed this process in the deep ocean it has been seen in the upper ocean[11]. Benthic fronts are potentially important because concentrations of tracer within and above the BBL depend in part on the frequency and efficiency of this process.

(c) Mixed layers detached from the seabed

Elevated mixed layers have been detected over an abyssal plain on the N.W. Atlantic[12]. Similar layers are seen in the temperature profile measurements made on both the Madeira Abyssal Plain and the Madeira Rise. The measurements made in the N.W. Atlantic show that these elevated layers can originate from a detachment of the benthic boundary layer from the seabed[12]. The modelling suggests that areas of detachment are created by the interaction of the bottom layer with eddies. Weak but persistent sinking of sea water in an anticyclonic eddy causes a warming of the bottom layer relative to the surrounding fluid and can lead to the formation of fronts. Buoyancy forces will cause the warmed mixed layer to detach from the bottom. According to the numerical models these areas of possible detachment occur over 10-25% of the total horizontal area of the bottom layer.

Another mechanism for detachment is possible when density surfaces intersect the seabed either on a sloping bottom or on a flat bottom when density surfaces are inclined to the horizontal. Mixed layer fluid can then escape from the seabed along its own density level, interleaving with the surrounding fluid[13].

Enhanced bottom layers have been seen in the lee of abyssal hills as well as elevated layers separating from the crest of a short ridge.

(d) Integrated effects of vertical mixing

Although several processes have been identified acting within and above the BBL to promote loss of tracer from the deepest parts of the water column, rates are not easily estimated even with the aid of numerical models. A quite different approach to the estimation of such a loss is to examine heat (or salt) budgets, as has been done in the western Atlantic[14,15]. In brief, the method is to measure the flux of heat and salt into a deep basin through a single narrow channel. If conditions within the basin are steady, the effects of the channel flux are balanced by vertical mixing within the basin: if the first is measured the second is thus inferred. According to a recent review of methods of estimating vertical (diapycnal) diffusion in the ocean[16], the method is powerful and the results robust.

Such an experiment was conducted under this contract in the N.E. Atlantic. As seen in figure 1 the Iberian and Madeira Abyssal Basins are separated by a ridge, the East Azores Fracture Zone, through which a narrow channel runs, location 2. Bathymetric surveys carried out on I.O.S. cruises have provided a detailed picture of this channel which has been named 'Discovery Gap', see Figure 9. The channel axis which lies in the direction 240-060°T is about 150km in length with a width of as much as 50km and as little as 10km. A sill lies at the SW end of the valley at a depth of 4675m about 500m above the plains to its north and south and about 500m below the crest of the valley walls.

Repeated observations made with a lowered recording thermometer in the 'Narrows'part of the channel (section BB[1] of Figure 9) reveals the colder water at the bottom and heaped against the S.E. side of the channel (Figure 10a). A variation in temperature and hence density across the channel implies a pressure gradient in the same direction, and hence flow down the channel. On sections beyond the two exit sills of the valley the cold water can be seen discharging (Figure 10b, c) into the deeper areas north of Discovery Gap.

Direct current measurements confirm and expand this picture. Eight neutrally buoyant floats ballasted for depths near 4700m suffered a N.E. displacement in directions between 030 and 080^{o} over a period of ten days with an average speed of 4 cm/s (Figure 11). Two floats ballasted at 4000m depth moved more slowly at about 2 cm/s revealing that the stronger flow was very near the bottom. An array of moorings (four across the Narrows section, and one on each of the exit sills) were deployed for a one-year duration and details of the measurements and the results are summarised in an earlier report[3].

Here the results of the current measurements are illustrated in Figure 12, from the Narrows and figure 13 from the exit sills. Persistent year-long flow through the channel is revealed at four of the six moorings, while at the other two interruptions at 2-10 day intervals are seen. The mean flows in the Narrows section when resolved in the down valley direction (060^{o}) are plotted as a function of height and position on Figure 14, superimposed on temperature contours also derived from the year-long records.

The flux of cold water through Discovery Gap can be estimated directly from the current meter measurements, see Figure 14. Water colder than a (potential) temperature of 2.05°C is discharged through the Narrows section and hence across the exits at a rate of $(0.21 \pm .04) \times 10^6 m^3 s^{-1}$ the uncertainties deriving mostly from interpolation of the currents between the moorings and extrapolation beyond them. If this water fills (part of) the bottom of the Iberian basin more or less continuously but the temperature everywhere remains steady, there must be a flux of heat into this water. The heat balance equation may then be written

$$\int_{\theta < \theta_F} (\theta_F - \theta)\delta V = \left(G/_{\rho}Cp + K_V \frac{\partial\theta}{\partial z}\right) A$$

which states that a flux of water δV entering the basin at a temperature θ is warmed to a temperature θ_F (here taken as 2.05°C) and the heat to provide this is supplied by the geothermal heat flux across the seabed and by turbulent (or diapycnal) diffusion across the constant temperature surface θ_F. A is the area of the surface and K_V is the mean (diapycnal) diffusion coefficient across it.

Inserting values

$G = 1.5\ \mu cal\ cm^{-2} s^{-1}$ [7], $\left(\frac{\partial\theta}{\partial z}\right)_{\theta_F} = 1.0 \times 10^{-4}\ {}^{\circ}Cm^{-1}$, and

$A = 1.2 \times 10^{11} m^2$ (these latter figures obtained from the author's data reports[18,19] and other historical measurements) yields

$$K_V = 2 \times 10^{-4} m^2 s^{-1} \text{ or } 2\ cm^2 s^{-1}$$

Very similar values have been obtained in the western Atlantic following the investigation of northgoing currents in the Vema channel[14] and on the Ceara rise[15]. The errors in our estimate of K_V are determined principally by uncertainty in the area A, which could be 30% larger or a factor of 2 smaller. Thus K_V lies in the range 1.5-4 $cm^2 s^{-1}$. There is also an uncertainty associated with the assumption of steadiness. The average transit time for cold water discharged from Discovery Gap to reach and cross the surface $\theta_F (= 2.05^\circ C)$ is $A.\overline{h}/V$ where $\overline{h}$ is the mean height of the θ_F surface above the seabed and $V = \int_{\theta \leqslant \theta_F} \delta V$ is the rate of discharge. Because $\overline{h} \approx 100m$ only, the transit time is determined as between 1 and 2.5 years depending on the value of A assumed. The assumption of steadiness then requires no substantial change in A within this period. Unfortunately our data at present is too sparse to illuminate this requirement. The consequence of these estimates for the dilution and spreading of a cloud of passive material are described in the final section of this paper on pages 31 et seq.

HORIZONTAL OR ISOPYCNAL MIXING

In contrast to mixing in the vertical, mixing in the horizontal is estimated more directly from the experimental measurements. At all three sites clusters of about 10 neutrally buoyant floats were deployed at heights between 100 and 400m above the seabed. (The floats, which were tracked acoustically, move with the water horizontally but do not follow its vertical displacements). The dispersion of the floats can be measured directly and hence the rate of mixing inferred. Because of the short duration of the experiments these estimates are quite unreliable[2]. An alternative method is proposed which utilises the long duration data from moored current meters. Dispersion studies have also been carried out using the numerical model[20]. The numerical experiments compliment the estimates of dispersion made from the observed data and extend the scales covered to 100-200km.

(a) Comparison of float and interpolated trajectories

On the Madeira abyssal plain the trajectory of two groups of four floats is compared in Figure 15 with the displacement of particles of water obtained by interpolating currents amongst a group of seven measurements made on moorings at distances between 5 and 50km. The method of optimum linear interpolation is employed, identical to that used in producing streamfunction maps of the flow, see Figures 2 and 3. Although small systematic differences are observed (of about 0.3 cm/s), which have their origin in current meter errors and differences in height between the observations, the agreement is satisfactory in view of the small currents (1 to 2 cm/s only).

On the Madeira rise a similar comparison has been made. Figure 16 shows the trajectory of six floats over a five day period along with interpolated trajectories calculated from the current meter data. The agreement is excellent, principally because of the swifter flows (about 5 cm/s). These experiences, together with other oceanic[21] and atmospheric[22] applications, confirm the power of the optimum linear interpolation method.

(b) Dispersion WITHIN eddies

The spread of a cloud of passive tracer can be determined from the rate of dispersion of pairs of particles with different initial separations. By the construction of interpolated trajectories for many pairs of particles the statistics of their behaviour has been determined. Figure 17 shows how the root mean square separation of a pair of particles increases with time on the Madeira Abyssal Plain. On average squared separation doubles in about 15 days a result which is independent of initial separations between 1 and 20km^2.

Similar calculations have been made from the data on the Madeira Rise, see Figure 18. In contrast these measurements show a doubling time for squared separation of about 2 days only. The large difference between this estimate and that from the Madeira Abyssal Plain is probably due to the larger gradients in flow at the former site. If L is a measure of eddy size and U a measure of its swirl speed then squared separation doubling time T is supposed[2] given by

$$T = C\,\frac{L}{U}$$

where C is a constant. Eddies on the Madeira Rise are smaller than those on the Plain by a factor of 2 and their flow speeds are 3 times stronger. The combination, a factor of 6 account for the majority of the difference in doubling time.

In Table 5 estimates of two particle dispersion from both the atmosphere and ocean are assembled. In the atmosphere both the results of trajectory calculations[23], of the same nature as those described in this report, and constant level balloon measurements[24] have been examined and the constant of the above equation determined as approximately 2. In the ocean the interpolated trajectory calculations and the dispersion of neutrally buoyant floats[25] (not those reported here) give a constant of similar value. There is more scatter in the oceanic estimates, in part from the smaller data sample. It is seen that for particle separations which are smaller than the scale of an eddy, atmospheric and oceanic eddies disperse particles in much the same manner and as described by the equation above.

(c) Dispersion BY eddies

Tracer mixing rates at scales larger than the eddy size are determined by the energy of the flow and the integral time scale[26]. This latter quantity I_L is given by

$$I_L = \int_0^\infty C(\tau)\,d\tau$$

Table 5 Two particle dispersion, atmosphere and ocean, separation <L

Source	Location	Method	Separation, km	Time,[1] days	Length Scale,[2] km	Fluctuation Velocity, m/s	Constant[3]
Kao & Al-Gain (1968)[23]	Atmosphere, 200mb N-hemisphere 500mb 30-60° 850mb	Trajectory calculation	600	1.7 2.3 2.9	1100 1100 750	16 13 10	2 2.3 2.9
Morel & Larcheveque (1974)[24]	Atmosphere, 200mb S-hemisphere	Constant level Balloons	80-1000	1.8±.2	1000	10	2
Saunders (1983)[2]	N E Atlantic ocean Abyssal Plain, 5300m	Trajectory calculation	1-20	15±5	35	.02	1
Saunders (1985)	N E Atlantic ocean Continental Rise 4500m	Trajectory calculation	1-10	2±.5	20	.05	0.5
Price[25]	N W Atlantic ocean LDE, 700m	Neutrally buoyant floats	30-100	11-13	50?	.10	2
Price[25]	N W Atlantic ocean 1300m	Neutrally buoyant floats	30-100	40-70	50?	.06	4-7

1. Mean time to double (initial) square of separation T
2. Distance to zero crossing of correlation of transverse velocity component L
3. In equation $T = c.^{L}/_{U}$

where $C(\tau)$ is the correlation function for a velocity component. This measurement is made following a fluid particle (called a Lagrangian measurement and hence the subscript L). Given current meter data from moorings, Eulerian measurements, an analagous quantity I_E can be determined where $C'(\tau)$ is the velocity correlation measured at fixed points. In the atmosphere it has been found that I_L and I_E are approximately equal for scales larger than eddies[27]. (I_L/I_E equal to 0.6 has been derived but the difference from unity will be ignored here). Hence in the ocean, for scales larger than eddies.

$$K_x,\ K_y \approx I_E(u'^2,\ v'^2)$$

where K_x is the diffusivity in the x direction and u'^2 is the variance of the velocity component in the same direction.

Table 6

Large-scale diffusivity in the abyssal N. E. Atlantic

Location	I_E (assumed = I_L) days		u'^2 cm^2s^{-2}	K_H $10^6 cm^2 s^{-1}$
Madeira Abyssal Plain	7±2	E*	3.6	2.2
(100m above bottom)		N	1.7	1.0
Discovery Gap	2.7±.8	C	3.2	0.8
(10m above bottom)		L	8.8	2.0
Madeira Rise	10±3	C	13	11
(100m above bottom)		L	24	21

* E - East N - North
L - rotated to principal direction of bathymetry, Longslope
C - cross-slope

Table 6 shows the result of estimating the integral time scale and deducing the 'eddy' diffusivity at the three sites. The value of $10^6 cm^2 s^{-1}$ for the Madeira Abyssal Plain[2] was the first estimate of this important quantity at abyssal depth. The high value of $10^7 cm^2 s^{-1}$ on the Madeira Rise reflects the unexpectedly energetic motions found there.

Data from a number of long-term neutrally buoyant float experiments have also been used to determine the diffusivity: the integral time scale from these experiments is approximately constant with a value of 7 to 10 days[25]. The estimate is not distinguishable from those made at two of the sites investigated under this contract. The quite different value of the integral time scale within Discovery Gap probably reflects the effect of the steep and irregular topography there.

(d) Numerical experiments

The numerical model provides predictions of the flow within and immediately above the benthic boundary layer due to a field of mesoscale eddies[28]. The dispersion of a tracer is studied in two ways. The tracks of a number of particles placed both within the boundary layer and above it have been calculated. The second and complimentary method is to calculate what happens to a single cloud of tracer under such flow conditions, see Figures 19,20.

The flow field for the experiments reported here has a r.m.s. velocity of 4 cms^{-1}, a length scale L = 50km* and a Lagrangian integral timescale of about I_L = 10 days. The initial separation of particles is 16km. Initially the particle pair separation on average grows exponentially with time. The average square separation doubles in about 15 days for particles in the boundary layer and 18 days for

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* Footnote Both the velocity and length scale probably exceed values representative of the N. E. Atlantic.

those outside. The constant in Table 5 is then C = 1 and 1.3 respectively, comparing well with the observations. For longer times the rate of increase of the average particle pair separation becomes constant and the diffusion process can be characterised by a constant eddy diffusivity K_H. This happens when the average particle separation is 70km or 1.4L in the boundary layer and 100km or 2L above the boundary layer. The diffusivities within the boundary layer and above are 5×10^6 cm s^{-1} and 8×10^6 cm^2 s^{-1} respectively which compare favourably with the estimate $K_H \simeq I_L u'^2 = 10^7$ cm s^{-1}.

The average time a particle will remain in the bottom layer is estimated by noting the time a particle enters a region of detachment predicted by the model[20]. This time is found to be on average 2 years, within which time about two thirds of the tracer will have escaped out of the bottom layer.

The numerical experiments of the dispersal of a cloud of tracer produced values of the diffusion rates[20] similar to those estimated from the ocean measurements. Initially the cloud will be pulled out by the eddies and become 'streaky' in appearance, see Figures 19, 20. At later times the cloud will fold back on itself and small scale diffusion processes will smooth out the large gradients in concentration. The cloud will become diffuse. Using the theoretical estimates of Garrett[29], on the Madeira Abyssal Plain the cloud will become well mixed after about 1 year. Numerical work is in rough agreement with this time[30].

ABYSSAL MIXING OF A PASSIVE TRACER

The estimates of horizontal (isopycnal) and vertical (diapycnal) mixing permit us to determine the spread of a passive tracer substance released from the seabed into the benthic boundary layer. The rapid diffusion rates deduced from observations of Radon 222[31] and theoretical work ensures that the tracer mixes relatively homogeneously within the BBL, say to a height of 50m above bottom, in a period of only a few days[5,32]. Progressive dilution of the tracer follows as it is carried further and further from the release site by the fluctuating currents. The currents keep the tracer approximately on a surface of the same density, an isopycnal surface. In the abyssal N. E. Atlantic such density surfaces correspond very closely to surfaces of constant potential temperature[33] which ascend equatorward at a rate of approximately 1km in 3000km[34]. For example, the surface of potential temperature 2.0°C intersects the bottom near 36N and at 10N its depth is only 4000m, over 1km above the bottom. Similarly the 2.1°C surface which intersects the sea floor near 43N is at a depth of 3700m near 10^0N. On such surfaces isopycnal mixing will disperse a tracer.

On the Madeira Abyssal Plain the squared separation of two particles doubles in an average period of 15 days. An exponential growth of separation results[24], namely

$$R^2 = R_0{}^2 \exp \alpha t$$

where $\alpha = {}^1/_{20}$ day^{-1} (since $\exp {}^{15}/_{20} \approx 2$). The above equation is valid so long as $R < L$, the scale of the eddy.

Given an initial separation of 0.5km, spreading to 20km takes place on average in approximately one half a year. Thus the concentration of tracer in a cloud of such material is diluted by a factor of $({}^{20}/_{0.5})^2$ or 1600 in this period.

At scales larger than the eddy size, the dilution is described by a different diffusion law, namely

$$R^2 = R_0{}^2 + 4K_H\ (t-t_0)$$

For a diffusivity $K_H \approx 10^6 cm^2 s^{-1}$ and an initial separation of 50km the squared separation will double and hence average concentration halve in about 60 days. For intermediate separations, namely 20 - 50km, there is therefore a weak decrease in the dilution rate, the measurement of which requires techniques different from those employed by us to date.

As described in an earlier section of this report, a survey of near bottom currents between 20 and 50^0N in the N. E. Atlantic reveals that over large areas of abyssal plain the eddy kinetic energy is approximately 1-2 cm^2s^{-2} [4]. In this context the Madeira Abyssal plain measurements appear typical and hence the dispersion characteristics measured there will be supposed applicable over the whole basein. [Further measurements and analyses of existing records are being undertaken to test the validity of this assertion]. On the basin scale a diffusivity of $10^6 cm^2 s^{-1}$ implies that a tracer will take approximately 20 years to spread through a region of side 1000km. Very slow persistent currents of a few mm/s, not measurable from the present short records, might slightly reduce, say halve, this period.

Mixing from a density surface to a neighbouring one (diapycnal diffusion) is a much slower process as the measurements in Discovery Gap[3] and other work shows[16]. Using a diffusivity K_v of 1 cm^2s^{-1} the vertical or diapycnal growth of a layer of tracer can be calculated from the expression

$$h^2 = h_0{}^2 + 4\ K_v(t-t_0)$$

where h is the layer thickness. If tracer initially fills the benthic boundary layer to a depth of approximately 50m then in one half a year the layer thickness will increase to only 100m. Thus whilst a tracer cloud spreads from 0.5 to 20km horizontally its thickness increases only from 50 to 100m. Hence tracer dilution due to isopycnal and diapycnal mixing is effective in the ratio 1600:2 or approximately 1000 to 1. This anistropy persists. After 20 years when the basin of dimension 1000km is filled laterally the tracer layer thickness is only about 1km deep. In the N. E. Atlantic because of the Northward and downward slope of the constant density surfaces (1 in 3000 approximately) the tracer layer will also be slightly wedge shaped.

Since the tracer is supposed passive there has been no assessment of the interaction of tracer substance with the bottom: neither absorption onto nor resuspension from the sediments is considered. Neither is the scavenging effect of particles falling freely through the tracer cloud nor the (radioactive) decay of the tracer itself considered. The final caveat concerns the omission of the uptake by marine organisms and the effect of the food chain. A short review of these processes and of work currently in progress concerning them is found in the recent report of the Holliday Committee[35].

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FIGURE CAPTIONS

Fig. 1 Map of N.E. Atlantic showing the location of the three sites.

Fig. 2 Maps of the flow 10m above the bottom on the Madeira abyssal plain with current vectors at the 7 mooring sites superimposed. The interval is 10 days.

Fig. 3 Maps of the flow 100m above the bottom on the Madeira Rise with current vectors at the 6 mooring sites superimposed. The interval is 5 or 10 days.

Fig. 4 Variation of mean current (u, v, Φ) and turbulence ($\overline{uw}$, $\overline{vw}$, k) in the BBL as a function of height above bottom: modelled numerically.

Fig. 5 Current vectors at 600m (upper) and 10m (lower) above the seabed on the Madeira abyssal plain. Tides and Inertial components have been filtered out.

Fig. 6 Current variability measured on a mooring expressed as a function of frequency or period. Periods of 1 - 100 days exhibit large differences between the three sites.

Fig. 7 Snapshot of the height of the bottom mixed layer taken from a numerical modelling simulation.

Fig. 8 Near bottom isotherms showing the repeated crossing of a front in a 6 day period on the Madeira abyssal plain.

Fig.9 Bathymetric map of Discovery Gap, depths in 100m. Mooring locations and temperature profile sections are shown.

Fig. 10 Temperature measurements in Discovery Gap made from a lowered thermometer: (a) Upper, along section BB' of figure 9, (b) lower left, along section CC' and (c) lower right, along section DD'.

Fig. 11 Track of 'constant level' floats in Discovery Gap Narrows at depths near 4700m. Positions at one day intervals with date. Depths in 100m.

Fig. 12 Current vectors 10m above bottom show the persistence of the flow in Discovery Gap Narrows: from moorings 321-4.

Fig. 13 Temperature and current vectors 10m above bottom on the two exit sills of Discovery Gap: moorings 326, 327 located on Figure 9.

Fig. 14 The computation of the flux of cold water through Discovery Gap Narrows.

Fig. 15 Comparison between the observed track of two groups of four floats (solid) and trajectories from current meter measurements (dashed) on the Madeira abyssal plain. Time in days/hours 1981.

Fig. 16 (Left) Track of six constant level floats (300m above bottom).

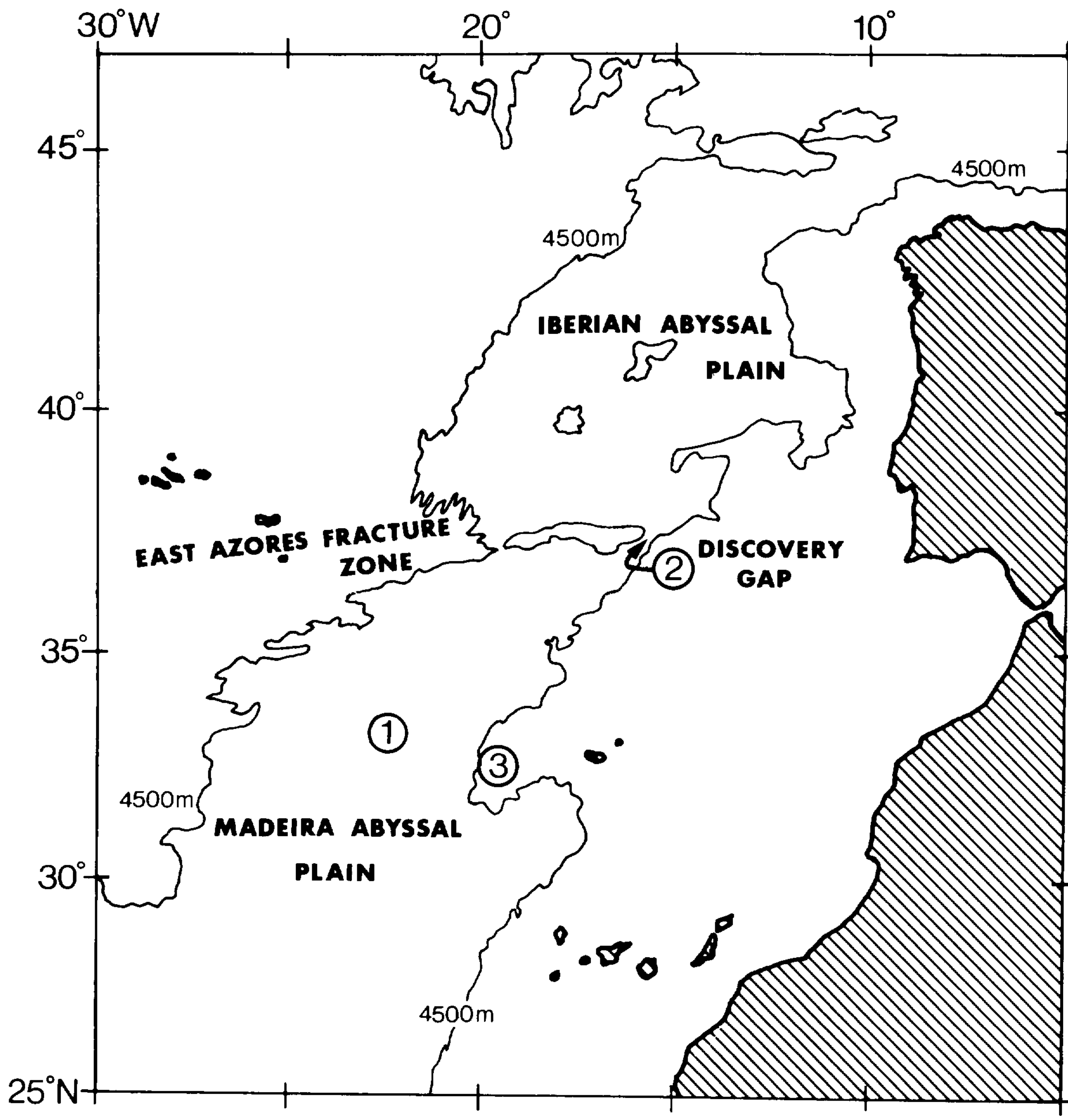
(Right) Trajectories interpolated amongst 6 current meter measurements (100m above bottom).

Fig. 17 The mean separation of pairs of particles calculated for flow 10m above the seabed on the Madeira abyssal plain.

Fig. 18 The mean separation of pairs of particles for flow 100m above the seabed on the Maderia rise.

Fig. 19 Evolution of a cloud of tracer 100m above the bottom from a numerical modelling simulation. The interval is 18 days and the box is 500km on a side.

Fig. 20 Evolution of a cloud of tracer within the BBL. Other details as for Figure 19.



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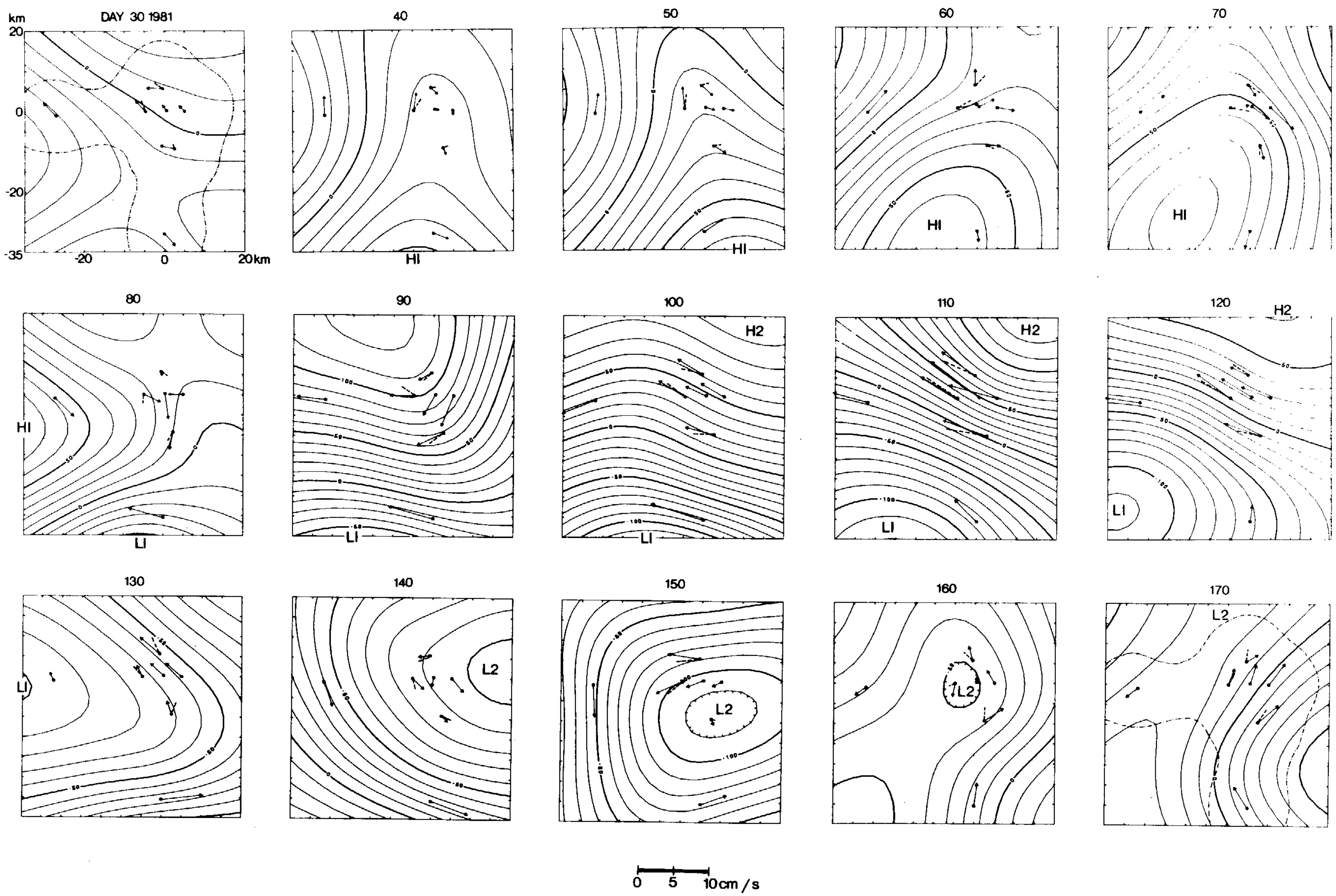

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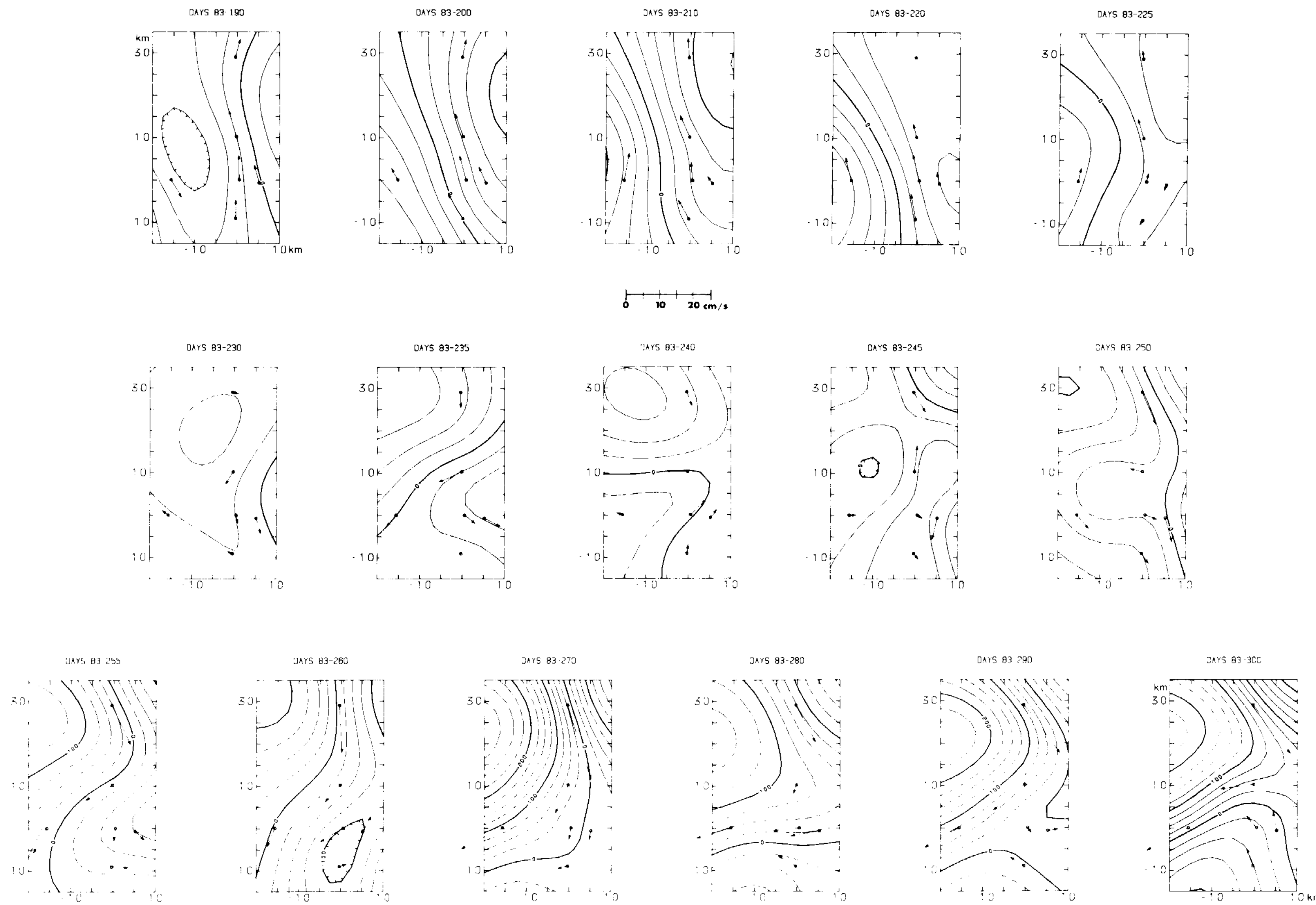

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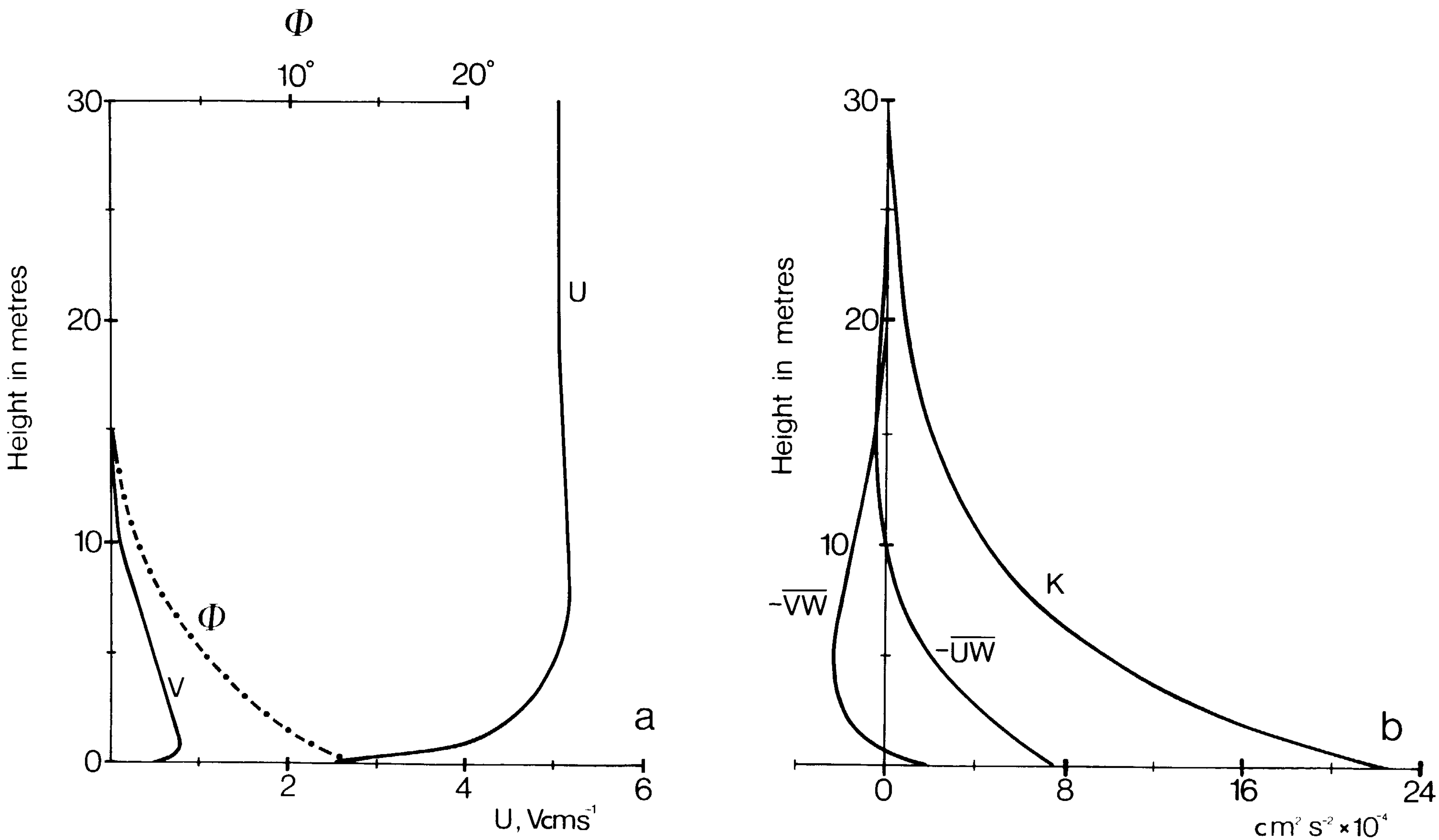

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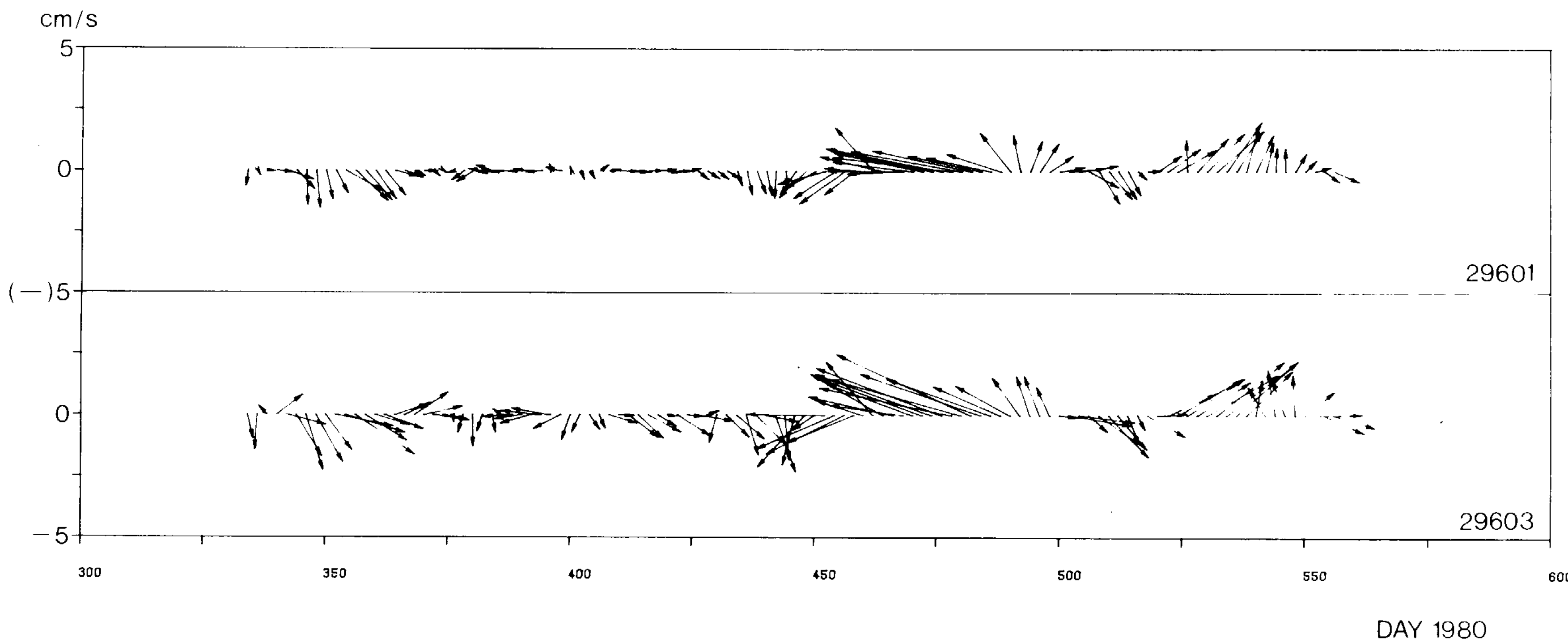

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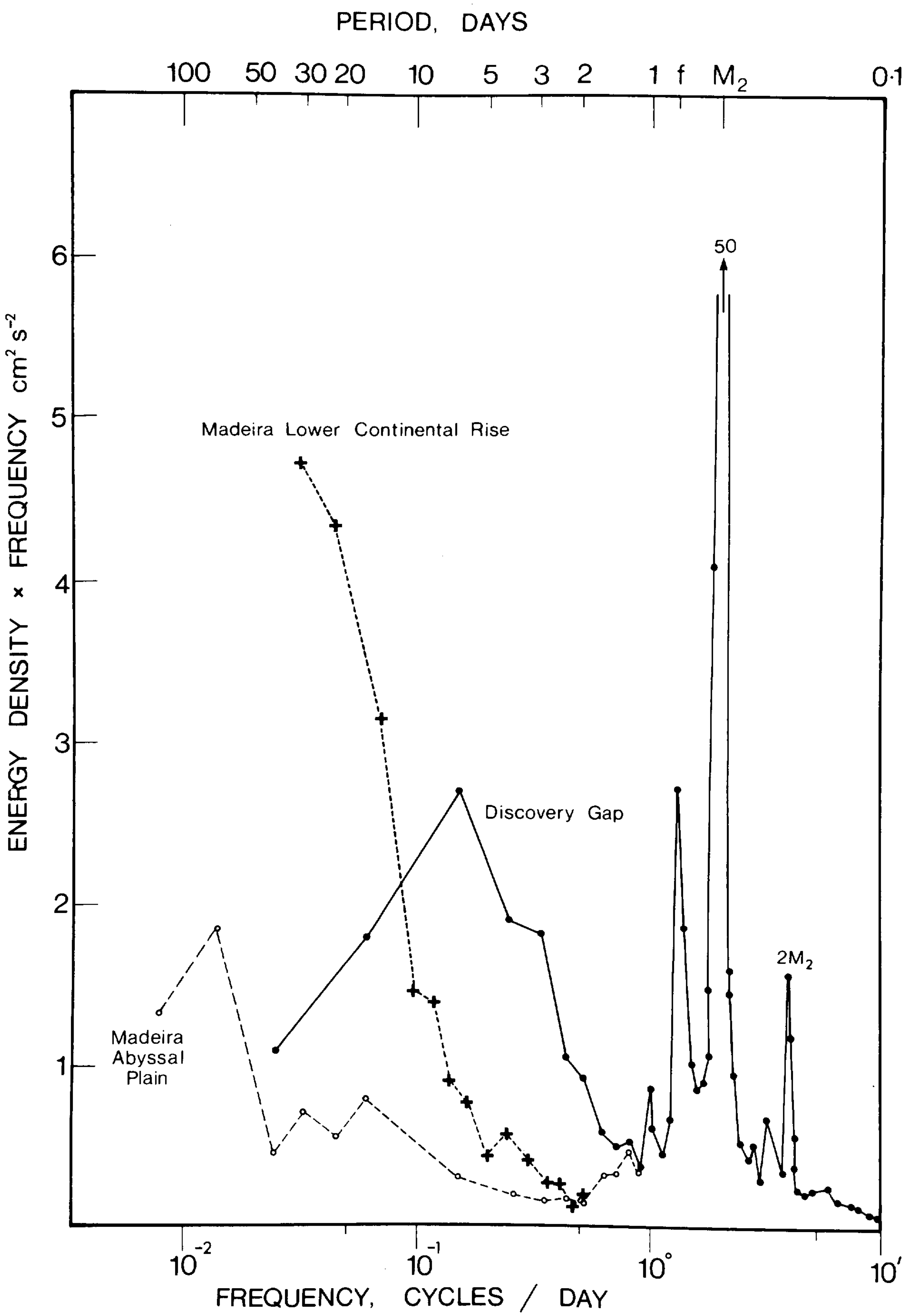

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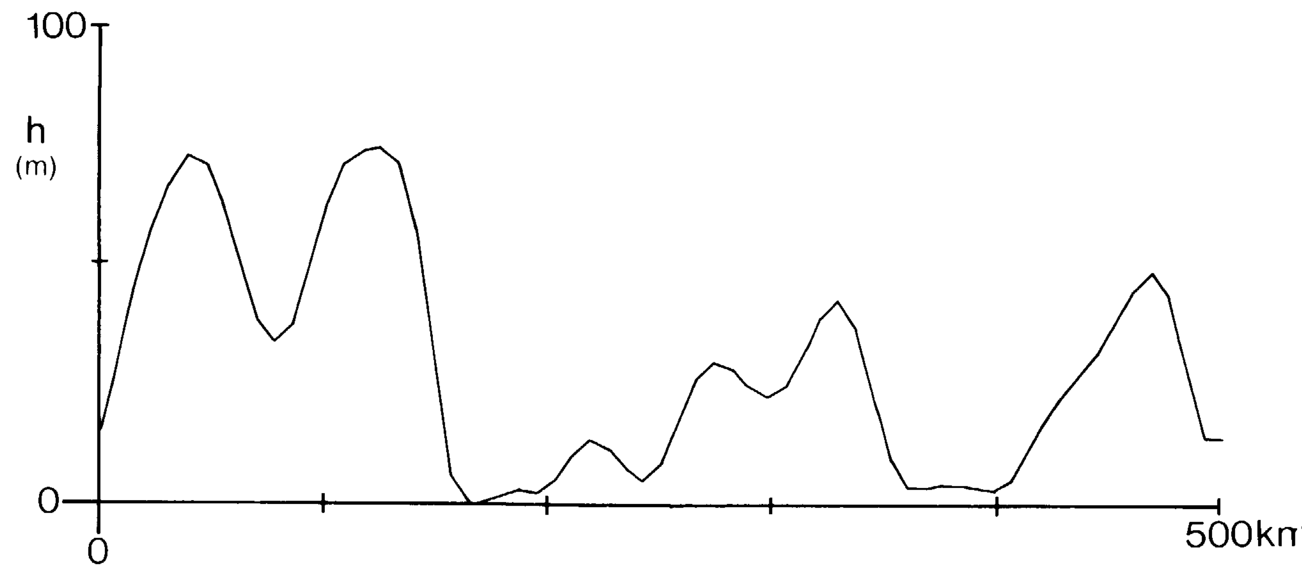

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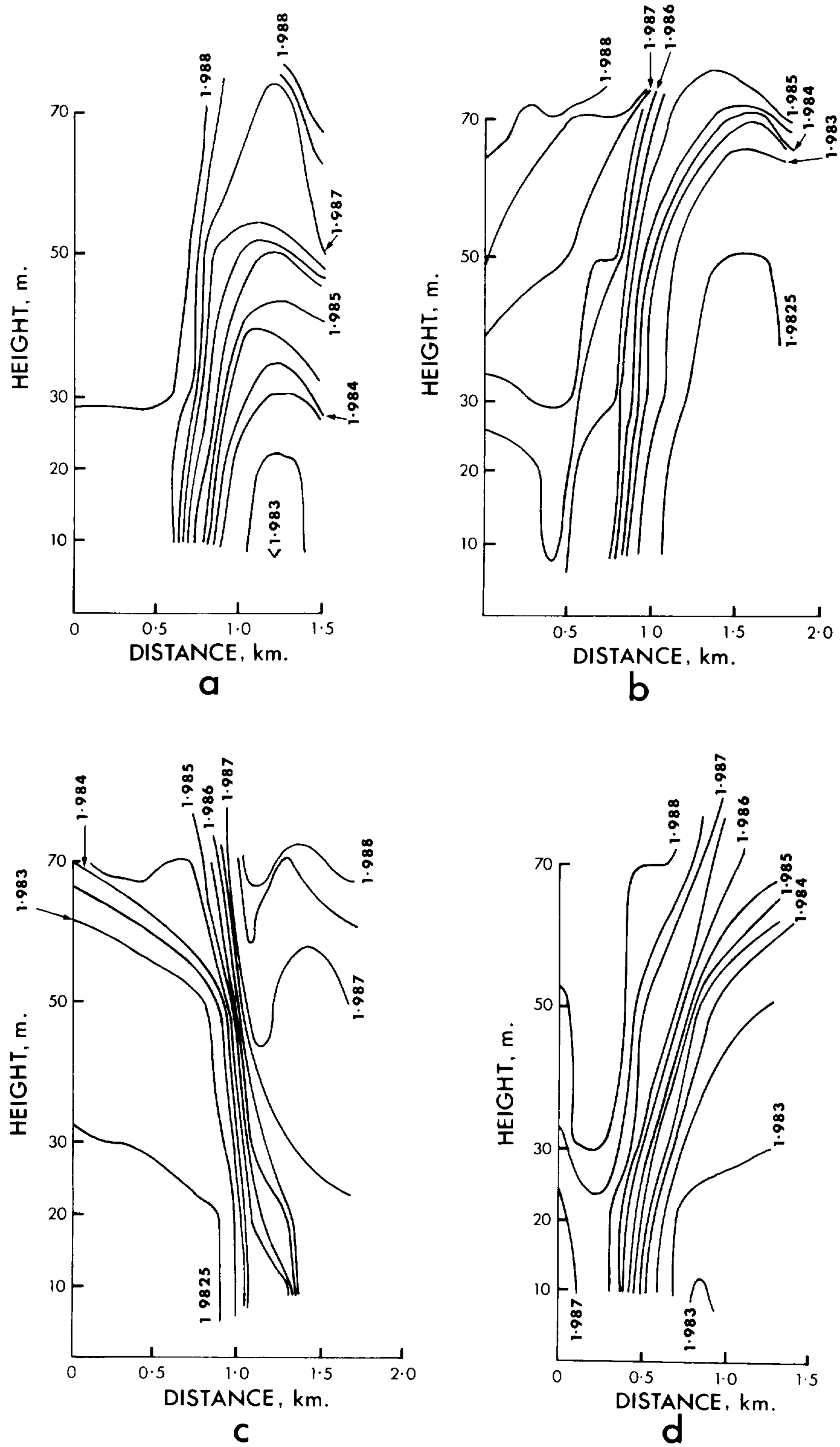

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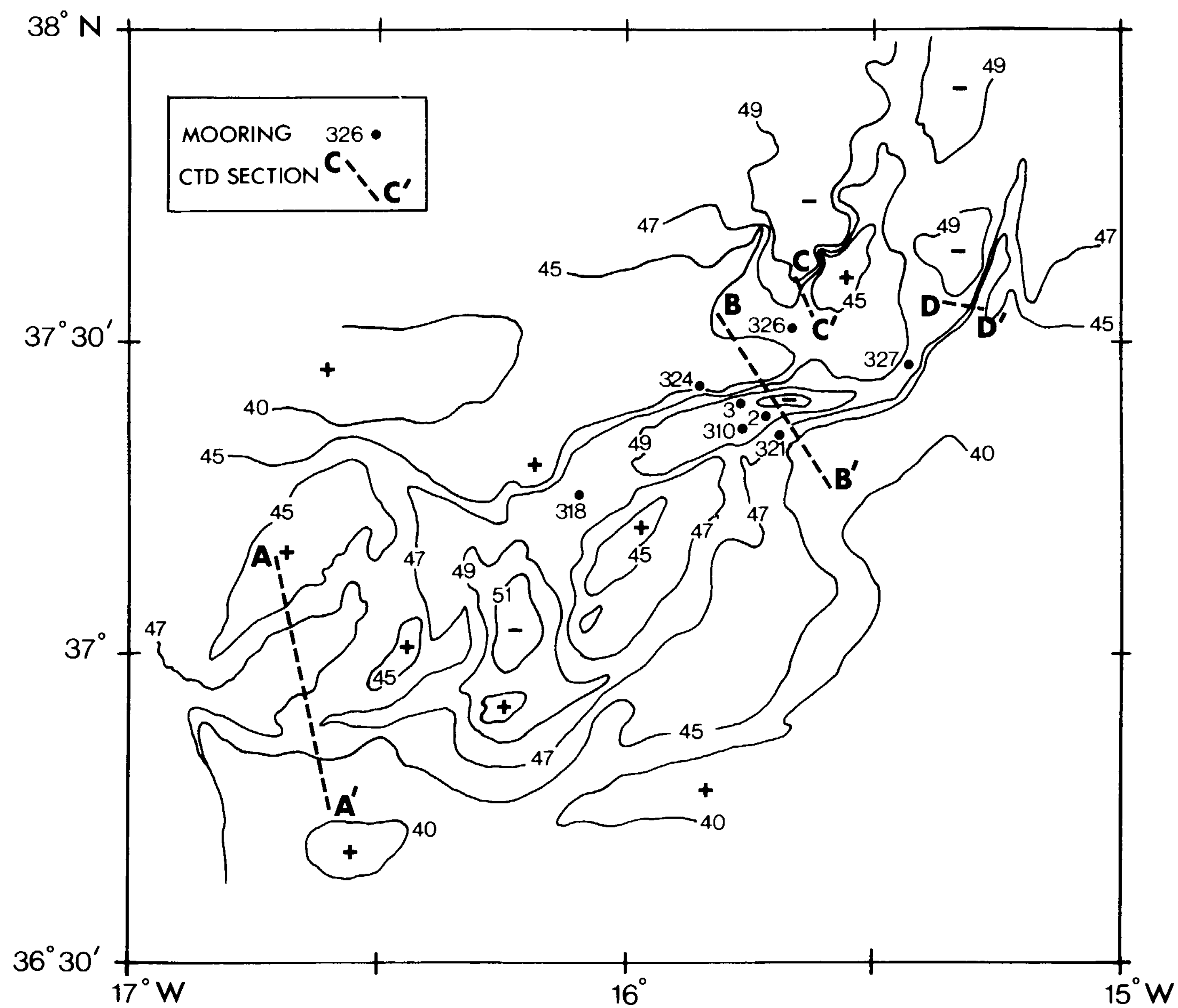

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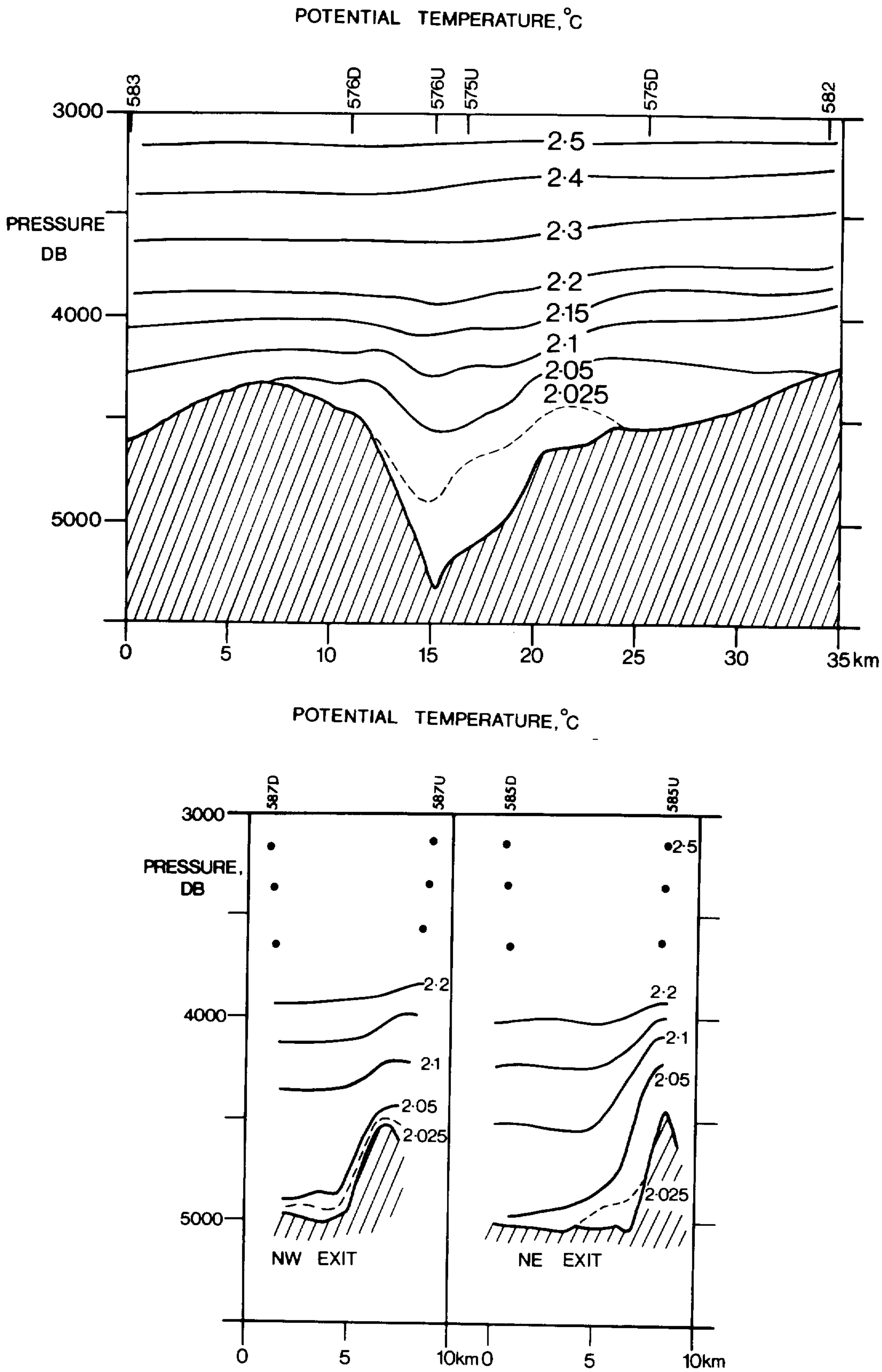

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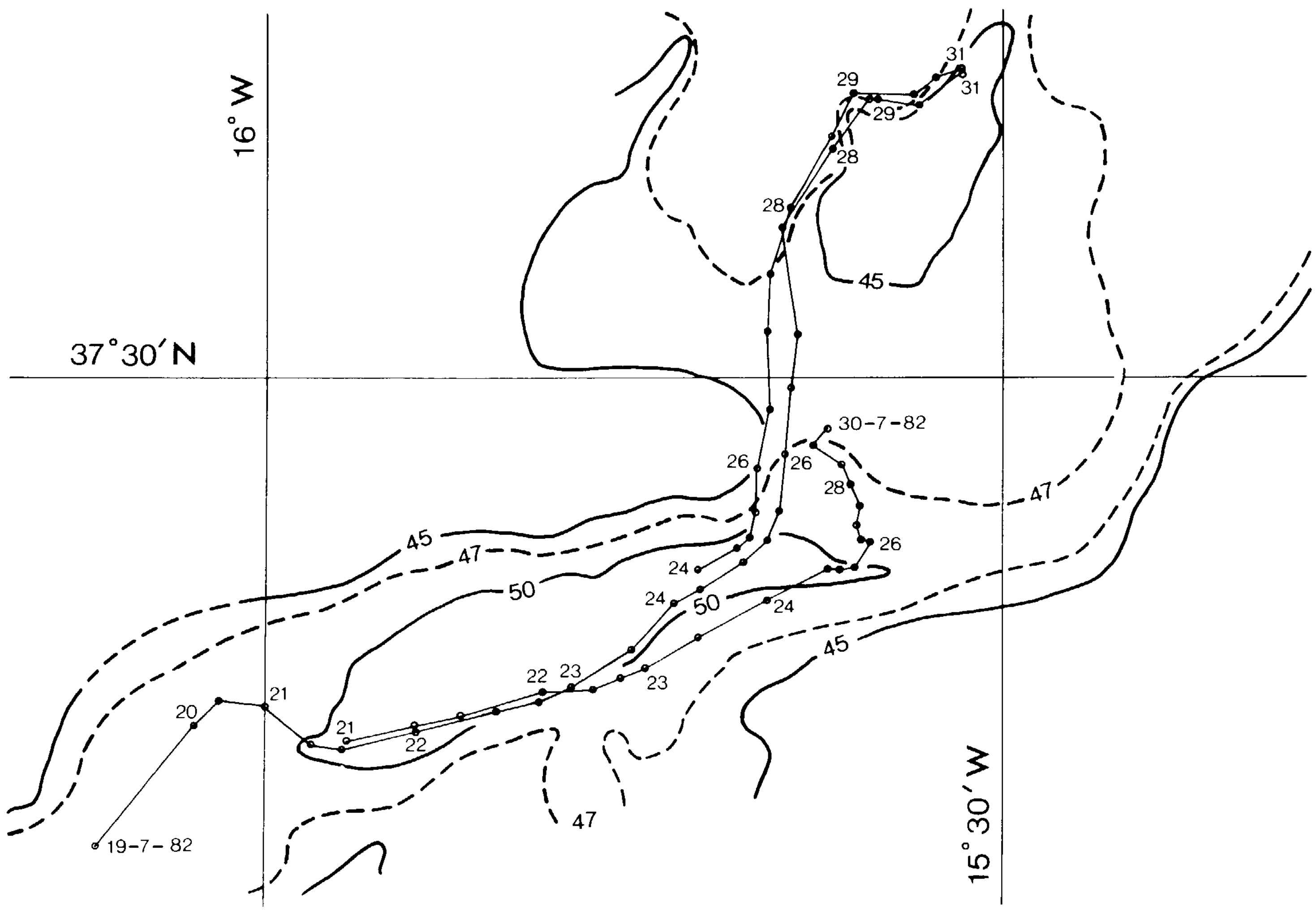

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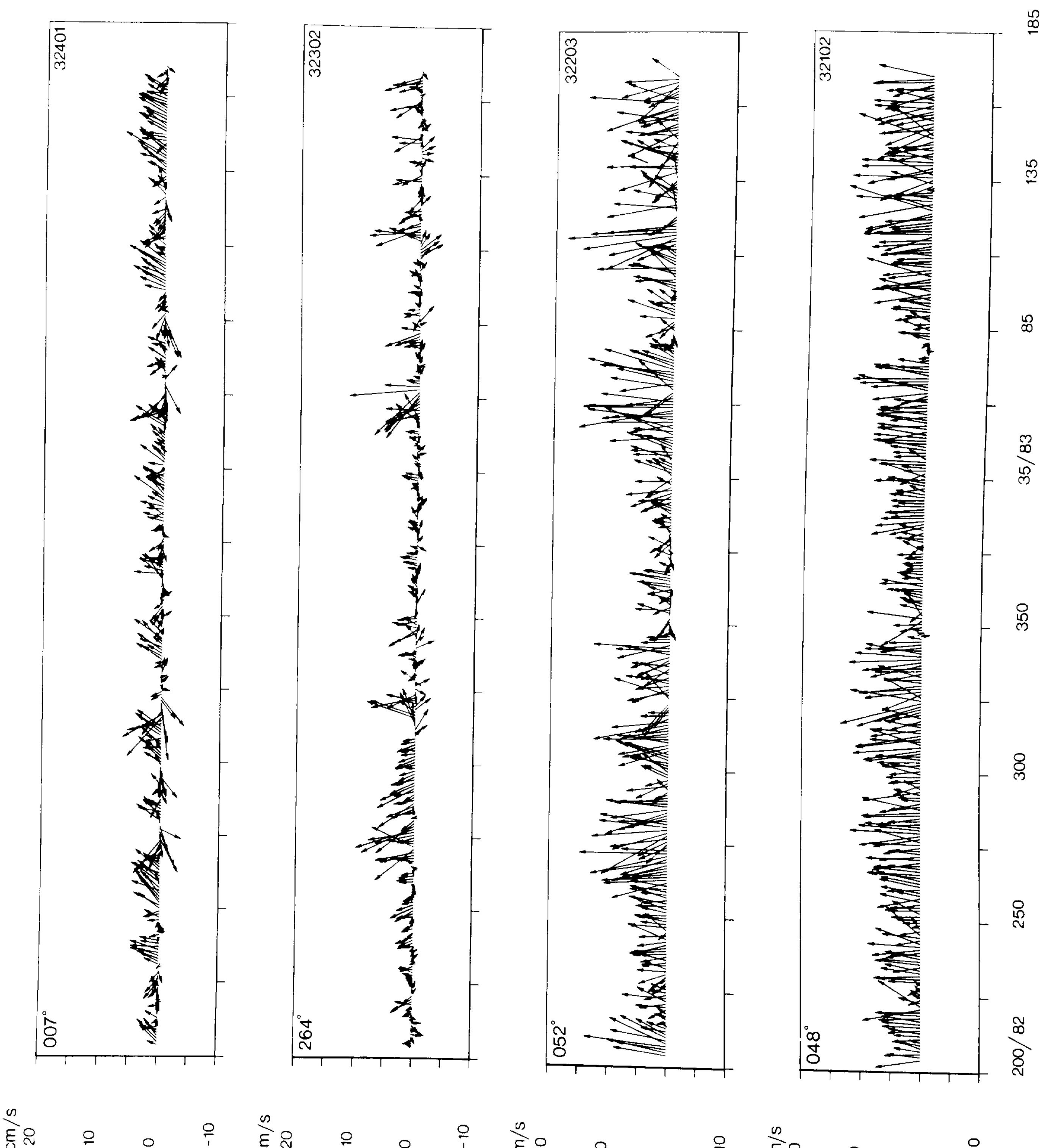

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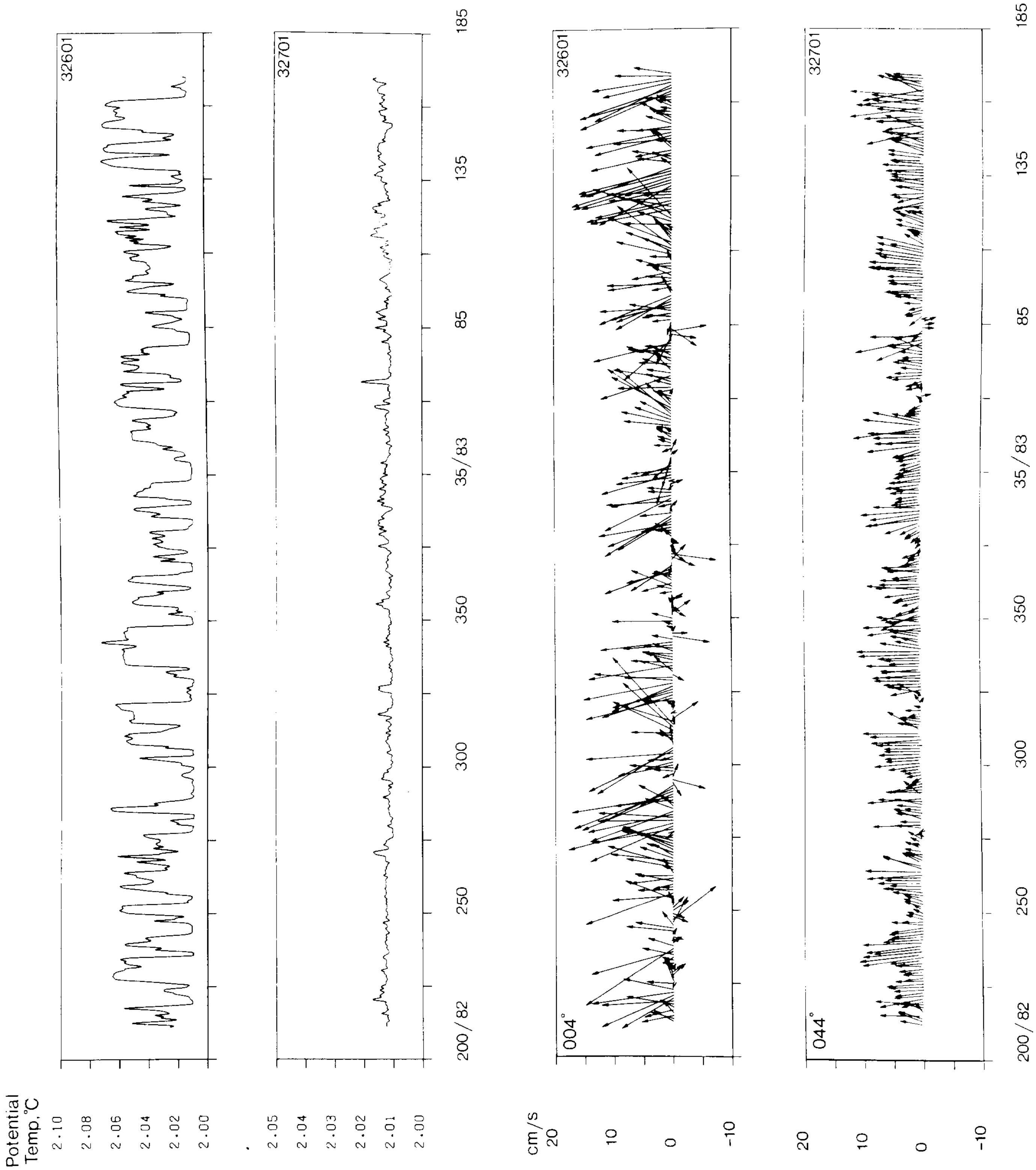

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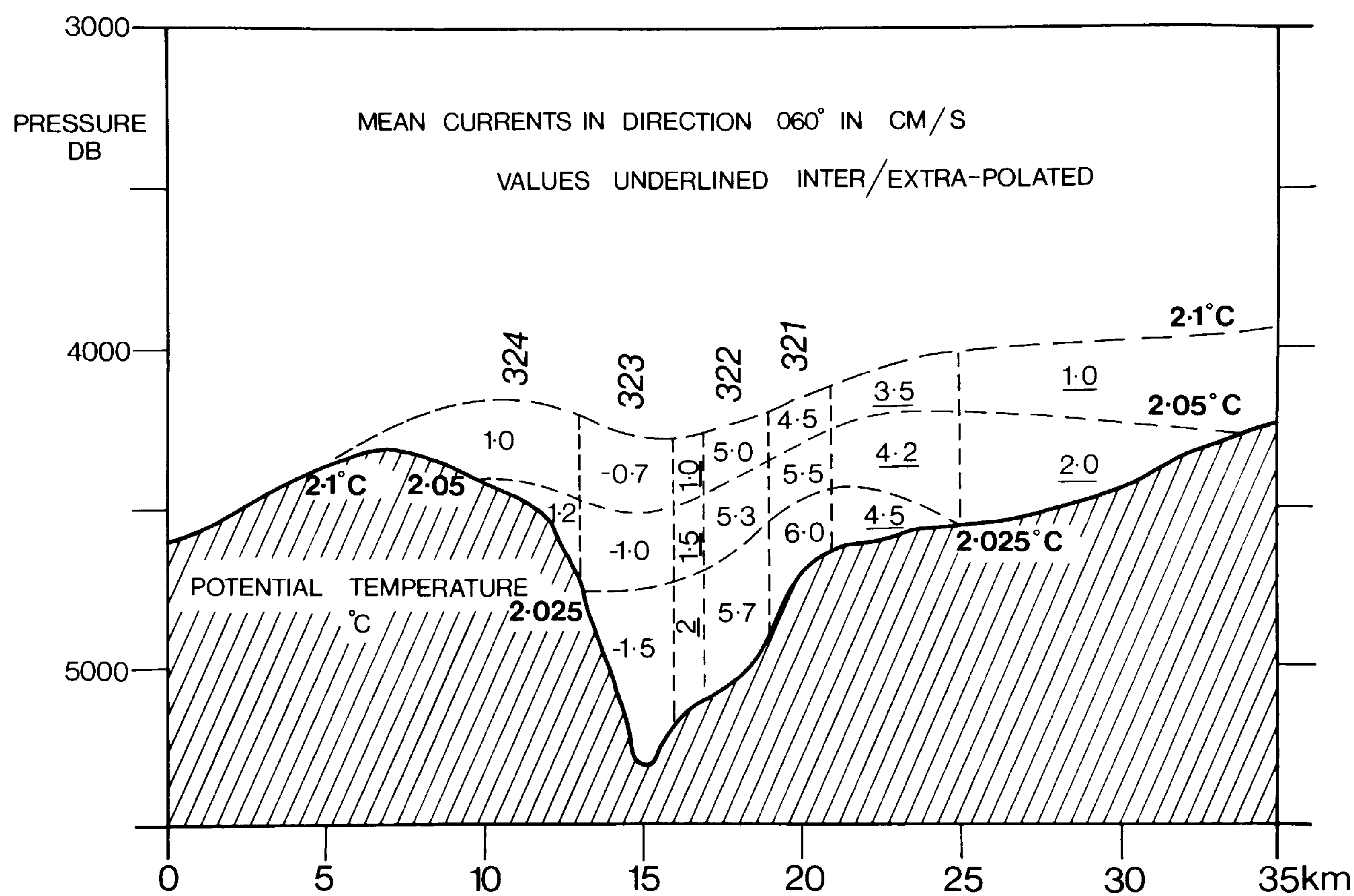

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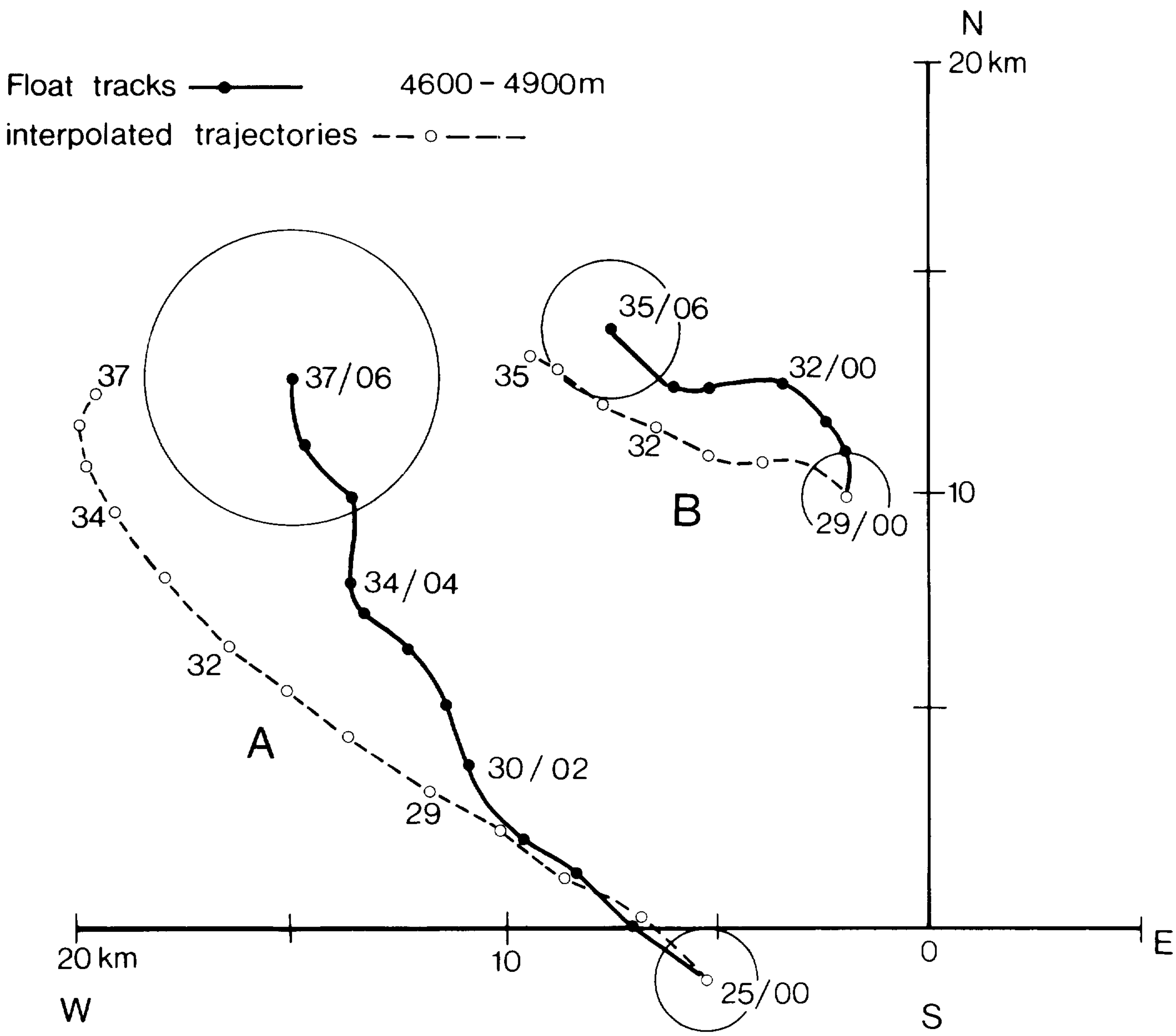

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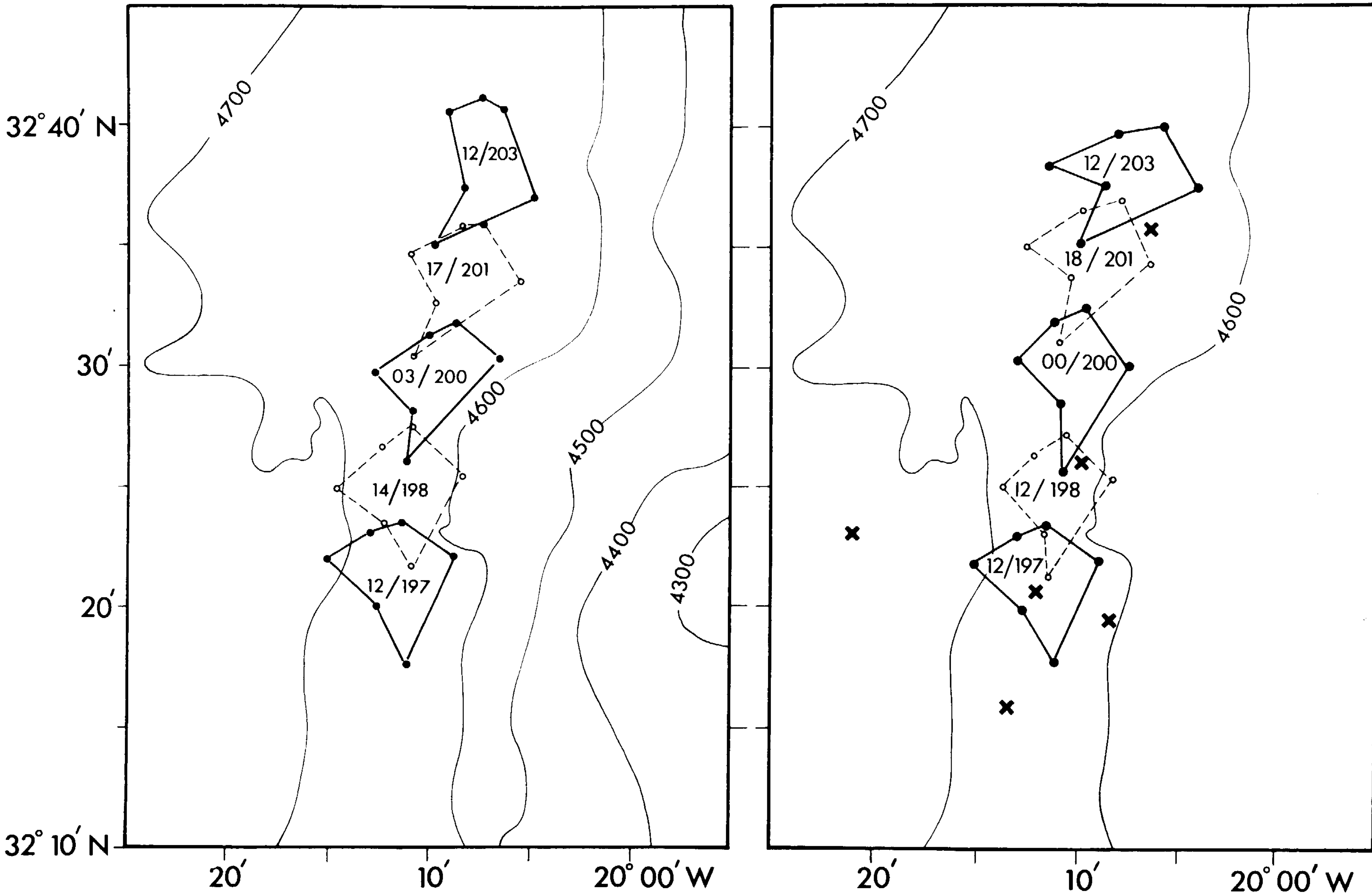

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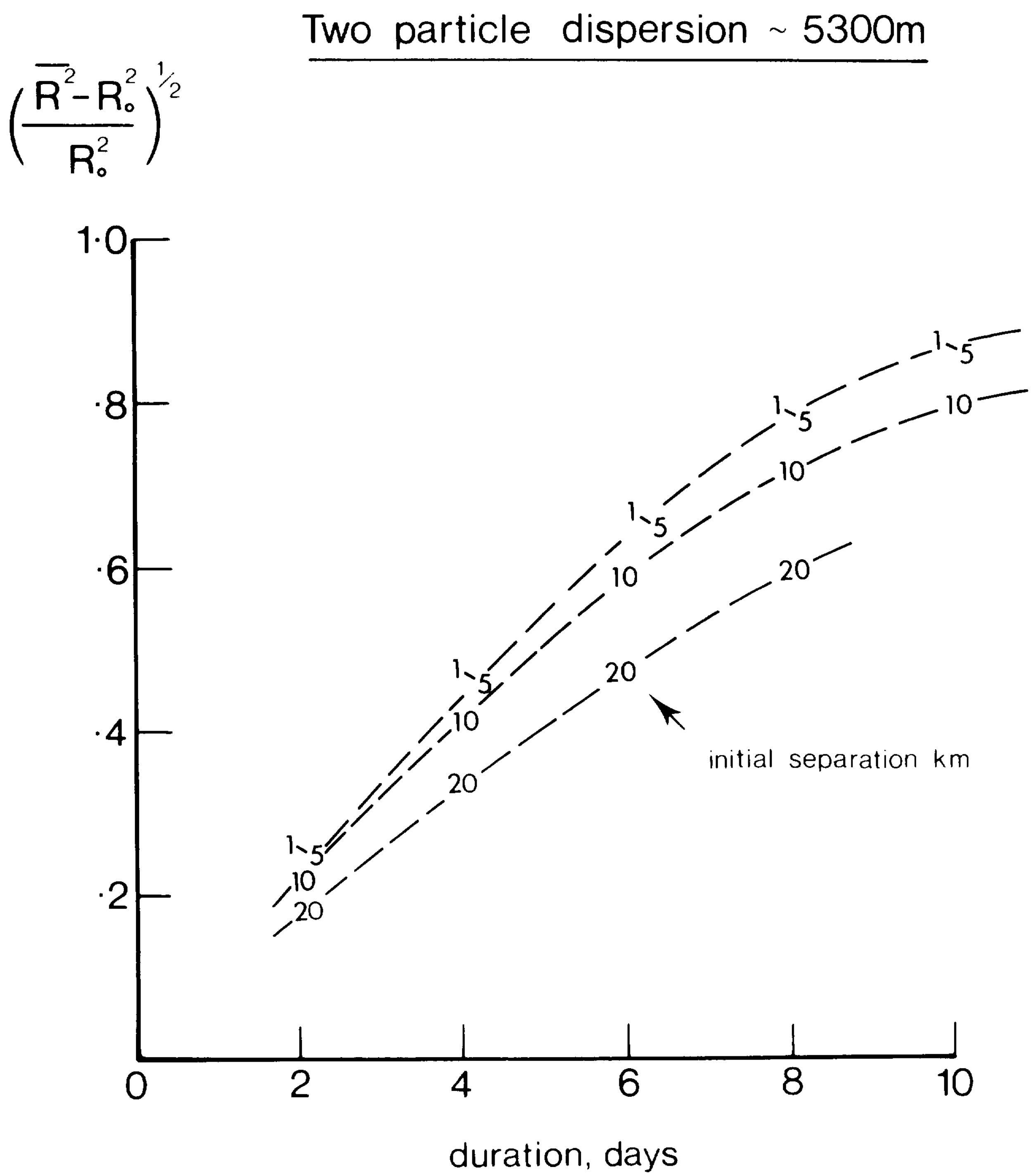

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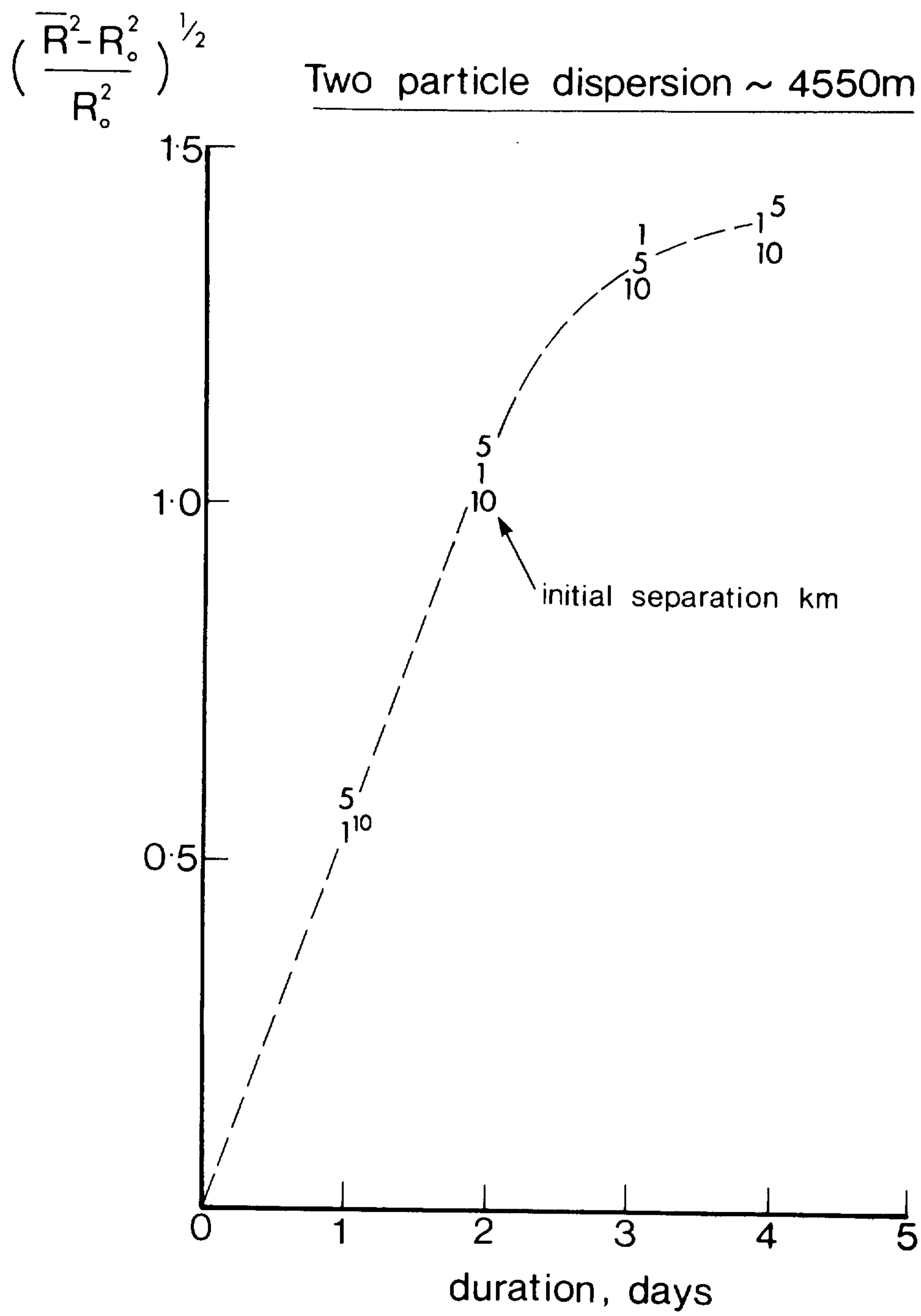

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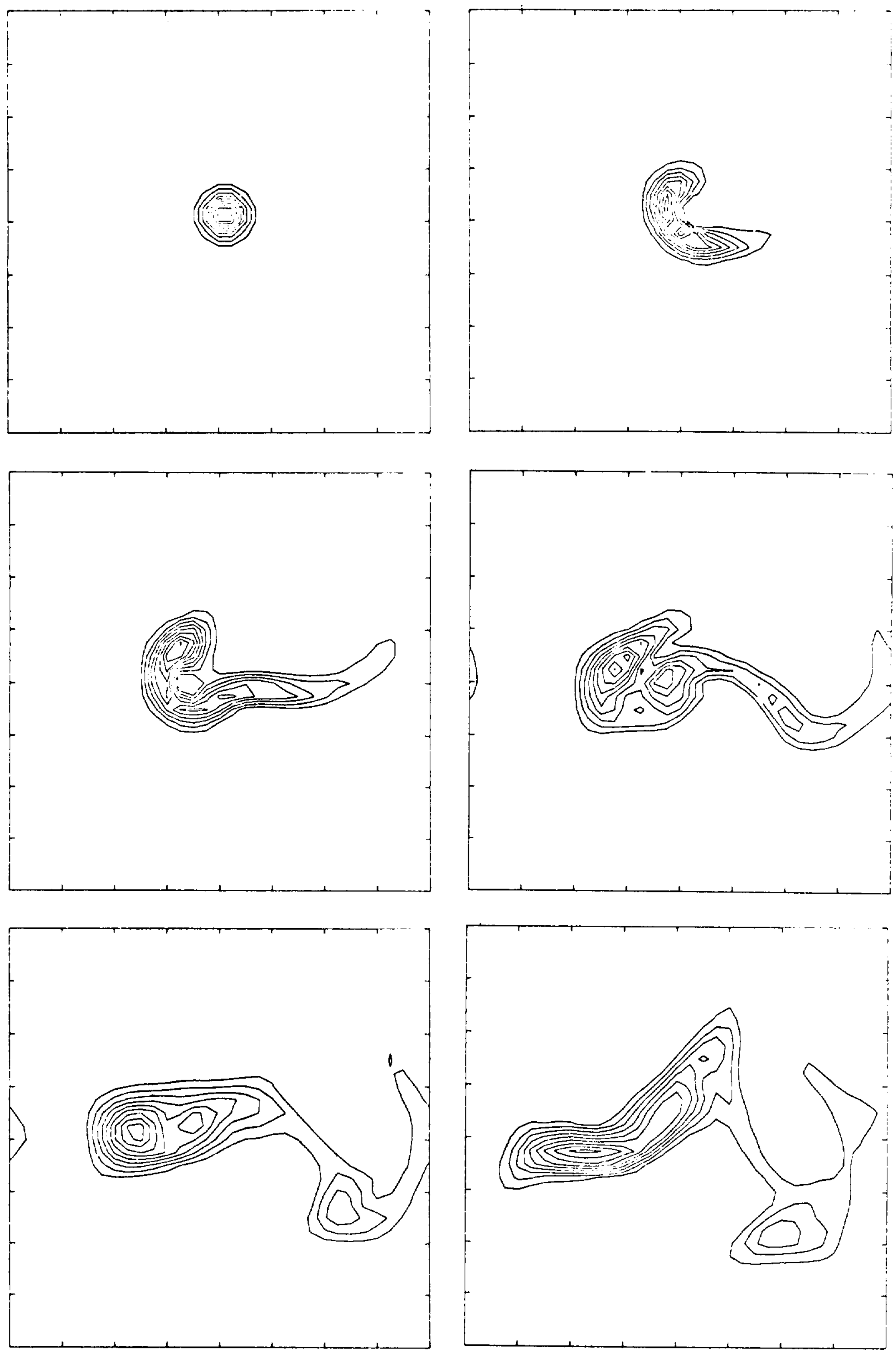

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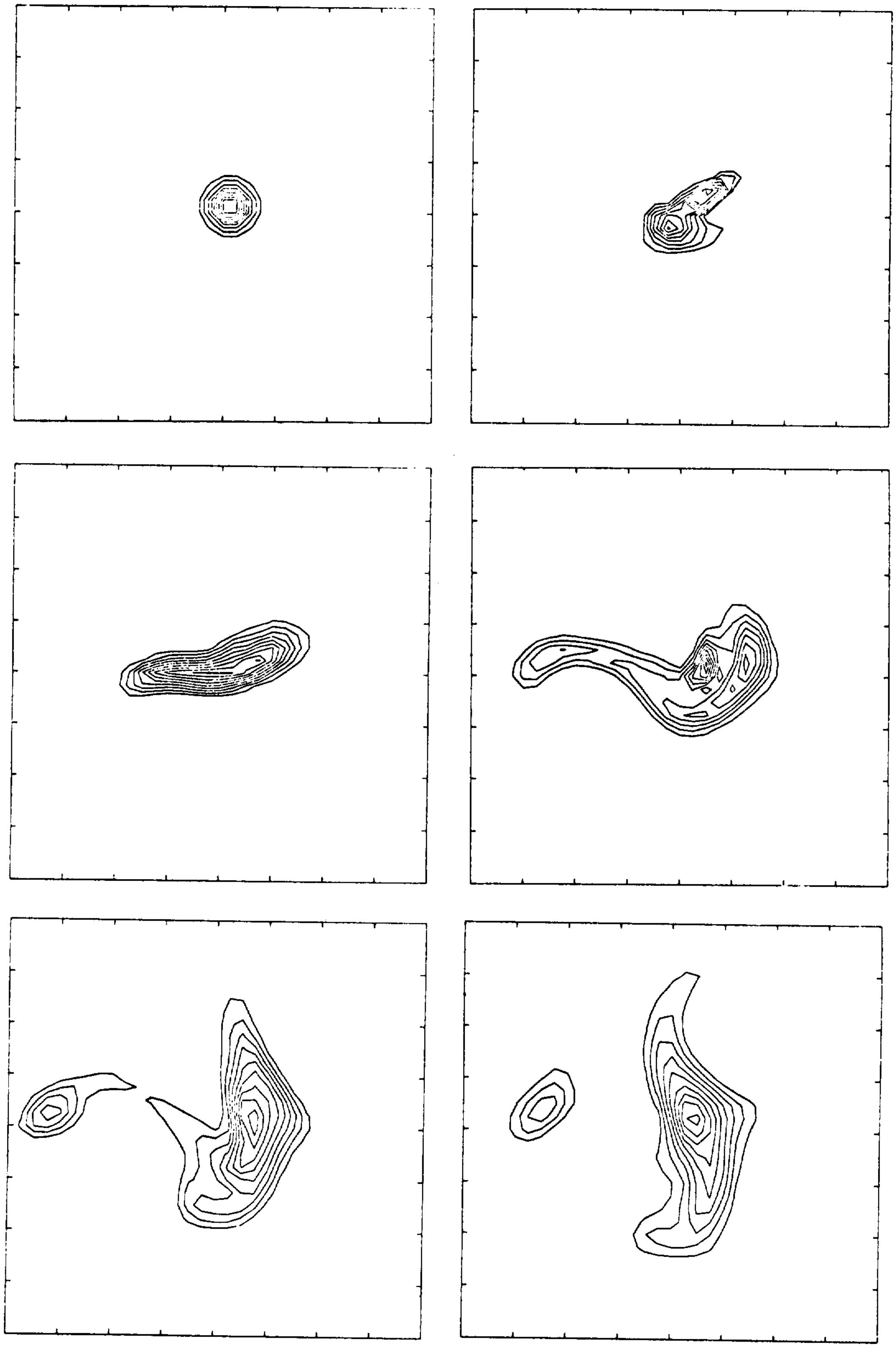

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