

M-type source material, unless the retention age of these components is much longer for M than that for P. Hence, by assuming that P and M in Fig. 2 represent the values for the source materials for the P-type and M-type, we can infer that the minimum values for the $(^3\text{He})_P/(^3\text{He})_M$ and the $(^{36}\text{Ar})_P/(^{36}\text{Ar})_M$ ratios are 5.5 and 57, respectively. This indicates that the P-type source material should contain larger amounts of primordial noble gas components compared with that of M-type. The apparent difference in the degree of retention between He and Ar in each source probably reflects their different mobilities.

To calculate the mixing rate between the two source materials, we need to know the isotopic ratios and relative concentrations for each endmember. Although P and M do not necessarily represent the endmembers, we assume that P and M approximate the values for each source material and that the relative concentrations of noble gases are represented by the minimum values stated above. In this case, we can get a result which corresponds to the minimum mixing rate for the M-type source material with the P-type source material. Even here, however, if a sample shows a $^3\text{He}/^4\text{He}$ ratio of 2×10^{-5} as in the case of Kilauea samples¹⁰, ~70% of the M-type source material might have mixed with 30% of the P-type source material. In general, the mixing rate of the M-type should be larger than this estimate, which means that Kilauea samples might have been produced by mixing more than double the amount of noble gas in M-type source material with that of P-type source material. These inferences give us constraints on the layered structure model of the mantle.

First, if the fertile mantle is located above the depleted mantle, such as in models *C-a* and *C-b*, an oceanic plate over it is considered to be the only source for the M-type source material. As long as we allow the mixing process to explain the observed data in Fig. 2, a mantle plume derived from the P-type source should melt more than double the M-type source material in the case for Kilauea samples. However, from the viewpoint of energy balance it is unlikely that a mantle plume could melt more than double the amount of lithosphere material.

Second, we observe large amounts of primordial noble gases in the fertile mantle. In models *B* and *C-a*, these are regarded as representing the primitive mantles, whereas in model *C-b* they are considered to be the result of an enrichment process such as mantle differentiation. In model *C-b*, the crust and the enriched mantle were formed through mantle differentiation producing the depleted mantle⁴. In this case, there is no clear distinction between the crust and the enriched mantle. To explain the difference in the observed $^3\text{He}/^4\text{He}$ ratios between the P-type and M-type sources by the enrichment process, a simple calculation indicates that such enrichment should have occurred before more than 2,000–2,500 Myr ago. As shown in Fig. 2, the continental crust seems to have lost most of its primordial noble gases. The high $^{40}\text{Ar}/^{36}\text{Ar}$ ratio for the M-type source also requires more than 60–70% of degassing of primordial components before more than 2,000 Myr ago from the mantle of M-type. Such degassing might have occurred through the combination of large-scale mass transport and the relatively large mobility of noble gases. Although we do not yet have sufficient information on the detailed mechanism of degassing from the Earth's interior, we consider that the mantle differentiation on global scale is probably accompanied by extensive degassing as long as the depleted mantle formed secondarily. Furthermore, if the P-type source was produced by the enrichment process and is located in the upper mantle without degassing, we cannot explain the separation of the P-type and the C-type materials as classified in Fig. 2. This implies the occurrence of rather different systems for these components. Hence, the enrichment process is not appropriate for explaining the abundant primordial noble gas contents in the P-type source material.

Third, mantle convection is likely to have caused large-scale mass transport. If it started early in the Earth's history and had

a speed similar to that of the present plate movement, more than 10 rotations of cells would be expected to have occurred in the layered mantle. Hence, taking into account both the mass transport by mantle convection and the relatively large mobility of light noble gases such as He, we consider it difficult for light noble gases to have been retained in a limited region of <100 km for more than several hundred million years. Helium at least would be degassed to the atmosphere unless it were retained at relatively deeper level such as the lower mantle.

Fourth, the $^3\text{He}/^4\text{He}$ ratios for MORB are reported to be relatively uniform over the Pacific, Atlantic and Indian Oceans¹¹. To produce such a uniformity in the $^3\text{He}/^4\text{He}$ ratio, a uniform $^3\text{He}/(\text{U}+\text{Th})$ ratio or some homogenized process such as a low-velocity zone⁸ is required in the mantle. He, and probably Ar, would be more easily homogenized through a partially melting part than through the solid part. On the other hand, samples of the P-type show more variation although they have systematically higher $^3\text{He}/^4\text{He}$ ratios than the MORB values. The variation can be most easily explained by model *B*, in which a mantle plume arises through the depleted mantle of M-type and it would be affected by mixing with the M-type source material in different degrees. Because the M-type material can extrude with no effect except for the surface materials, it will show much more uniformity in the isotopic ratios.

These conjectures lead us to conclude that noble gas isotope data are most compatible with model *B* out of the four shown in Fig. 1. In this case, a mantle plume at typical hotspot areas such as Hawaii and Iceland should arise from a relatively deeper mantle, though its depth cannot be determined from the present results. On the other hand, recent studies have revealed another type of hotspot such as Gough and Tristan da Cunha, where the samples show lower $^3\text{He}/^4\text{He}$ ratios than do the MORB values¹². These source materials would probably have been affected by a secondary process, such as contamination by crustal materials. Hence, only a typical mantle plume whose source is represented by a higher $^3\text{He}/^4\text{He}$ ratio compared with that of MORB probably arises from the relatively deeper part of the mantle which still retains primordial noble gases.

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Mantle eclogite and carbonate as sources of sodic carbonatites and alkalic magmas

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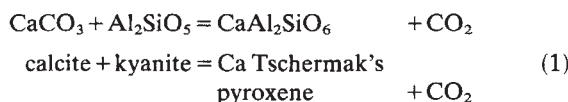
At mantle pressures, partial melts of peridotite+CO₂ are carbonate-rich, but contain insufficient sodium to form most carbonatites and associated silicate magmas. A possible source rock is eclogite, and extrapolations from experiments suggest that eclogite+CO₂ will coexist with melt along geotherms of low heat flow. Thermochemical calculations yield high carbon-

ate activities, and imply that the melt is carbonatitic. The melt can be modelled in the system jadeite–calcite–CO₂ and will, on emplacement, form mineral assemblages characteristic of a wide range of alkalic rocks. The tectonic settings of carbonatites are consistent with an origin as partial melts of eclogite + carbonate + CO₂ under areas of low surface heat flow, and suggest that carbonatite emplacement is not related to mantle hotspots.

At pressures above 25 kbar, partial melts in the system peridotite + CO₂ are rich in carbonate, and have been correlated with kimberlite or carbonatite magmas^{1,2}. However, peridotite does not contain enough sodium to allow formation of most carbonatites and their associated sodic mafic magmas (ijolite, melilitite), which contain sodic minerals or have produced sodic metasomatic aureoles³. The sodium is presumably derived from the mantle, as isotope studies of carbonatites indicate minimal crustal contamination^{4–7}. A possible source of sodium is jadeitic pyroxene (omphacite) in mantle eclogite (a widespread, although minor component of the upper mantle⁸), and we have considered phase relations among eclogite, carbonate, and CO₂ to determine whether carbonatites and their associated silicate rocks might be generated by partial melting of eclogite + carbonate + CO₂.

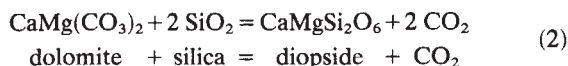
The mantle source of carbonatite melts presumably must contain significant carbonate, but a carbonate mineral (calcite) has been reported only once in mantle eclogite⁹. However, carbonation reactions involving omphacite (jadeite–diopside–Ca Tschermak's pyroxene solid solution), grossular–pyrope garnet, and minor eclogite minerals (kyanite, corundum¹⁰, or a silica polymorph^{11,12}) can be used to limit the activity of carbonate minerals in equilibrium with eclogite. Phase relations were calculated with the computer program EQUILI (Wall and E.J.E., unpublished), using free energy of CO₂ from Bottinga and Richet¹³, and data for solids phases from the literature^{14–16}.

Among the possible equilibria

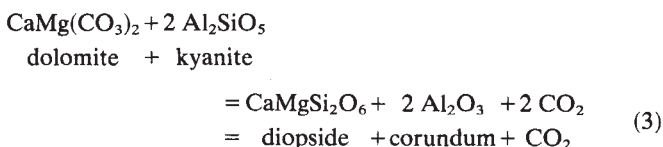


is the simplest and most widely applicable. Omphacite from mantle eclogite¹⁰ generally contains 5% Al(IV) and 60% Ca in the M2 site, so that the activity¹⁷ of Ca Tschermak's pyroxene (CaTs) in a typical mantle eclogite is estimated as $0.05 \times 0.60 = 0.03$. For such a pyroxene, Fig. 1a shows the ratio of calcite activity to carbon dioxide activity buffered by reaction (1). If kyanite is absent (as in most eclogites), the activity ratio isopleths of Fig. 1a represent lower limits, and calcite activity will be higher.

Similarly, carbonation reactions constrain the activity of dolomite in equilibrium with eclogite. The reaction



can be used to constrain the ratio $a(\text{Dol})/a^2(\text{CO}_2)$. Omphacitic pyroxene in mantle eclogite contains significant diopside component¹⁰, equivalent on average to $a(\text{Diop}) = 0.3$ using an ideal ionic activity model. Figure 1b shows $a(\text{Dol})/a^2(\text{CO}_2)$ buffered by such a pyroxene in reaction (2), with the quartz–coesite equilibrium¹⁸. For eclogites lacking quartz or coesite, the activity isopleths of Fig. 1b are minima, and dolomite activity will be higher. For kyanite–corundum eclogites¹⁰, activity ratio $a(\text{Dol})/a^2(\text{CO}_2)$ may be calculated from the reaction



Activity ratio isopleths for reaction (3) are shown in Fig. 1c; for eclogites containing diopside and kyanite, but no corun-

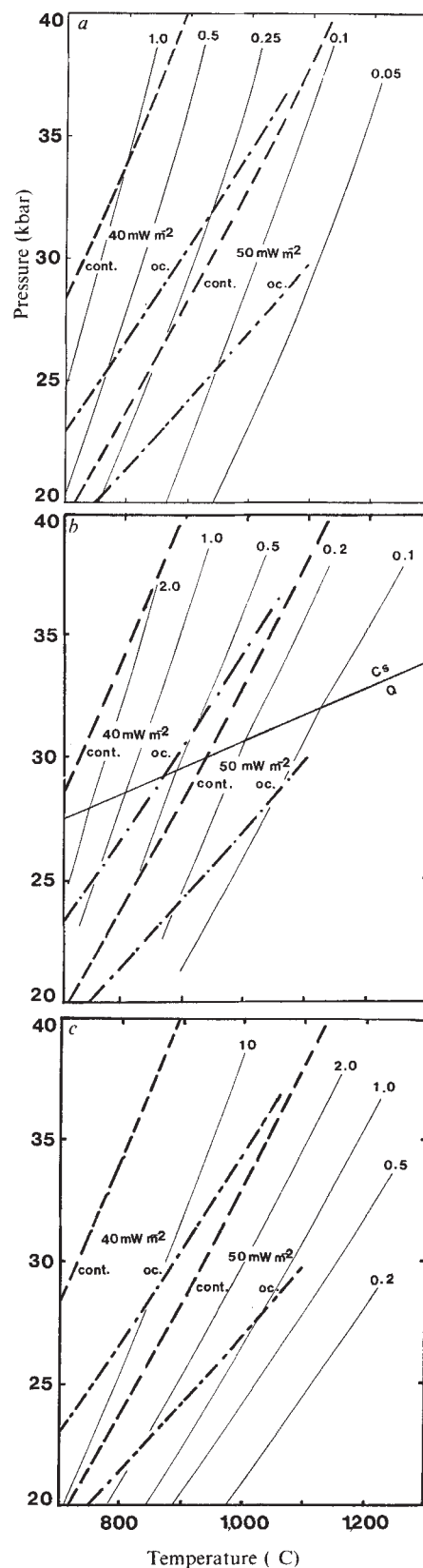


Fig. 1 a, Calculated ratios $a(\text{Cc})/a(\text{CO}_2)$ buffered by reaction (1) with $a(\text{CaTs}) = 0.03$. For kyanite-absent eclogites, plotted values are minima. b, Calculated ratios $a(\text{Dol})/a^2(\text{CO}_2)$ buffered by reaction (2) with $a(\text{Diop}) = 0.3$. For eclogites without quartz or coesite, plotted values are minima. c, Calculated ratios $a(\text{Dol})/a^2(\text{CO}_2)$ buffered by reaction (3) with $a(\text{Diop}) = 0.3$. For eclogites with kyanite but without corundum, plotted values are maxima. Calculated oceanic (oc.) and continental (cont.) geotherms and their surface heat flow are shown as dash-dot lines and dashed lines respectively.

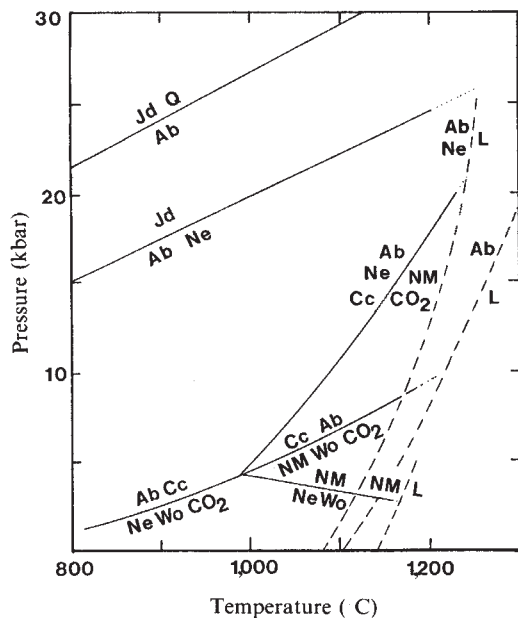


Fig. 2 Calculated phase relations in $\text{NaAlSi}_3\text{O}_4\text{-CaO-SiO}_2\text{-CO}_2$. Melting relations are for the vapour-absent sub-system¹⁷, as solidi for the vapour-saturated system are not known. Metastable reactions are dotted. Formulae and abbreviations for minerals in the system $\text{NaAlSi}_3\text{O}_4\text{-CaO-SiO}_2\text{-CO}_2$ are—calcite: Cc, CaCO_3 ; wollastonite: Wo, CaSiO_3 ; albite: Ab, $\text{NaAlSi}_3\text{O}_8$; jadeite: Jd, $\text{NaAlSi}_2\text{O}_6$; nepheline: Ne, $\text{NaAlSi}_3\text{O}_4$; soda-melilite: NM, $\text{CaNaAlSi}_2\text{O}_7$; quartz: Q, SiO_2 ; coesite: Cs, SiO_2 .

dum, the isopleths of Fig. 1c represent upper limits on $a(\text{Dol})/a^2(\text{CO}_2)$.

Essential to application of these equilibria is knowledge of CO_2 fugacities within the mantle. We assume that CO_2 pressure is a significant fraction of load pressure, because fluid inclusions from mantle rocks¹⁹⁻²¹ are generally rich in CO_2 , and experimental studies of carbonatite genesis yield results compatible with significant CO_2 in the source region for carbonatites. For pure CO_2 gas, the isopleths of Fig. 1 show that carbonate activities are significant only below 1,000 °C in the pressure range shown.

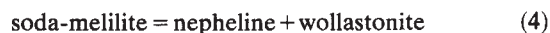
Carbonation reactions in the system $\text{CaO-Al}_2\text{O}_3\text{-SiO}_2\text{-CO}_2$ involving the phases (or components) calcite, grossular, Ca-Tschermak's pyroxene, kyanite, corundum, or quartz-coesite can be used as buffers of CO_2 activity. For the garnet and pyroxene of natural eclogite, calcite activities are similar to those of Fig. 1a. Similarly, in the system containing MgO (including the above minerals plus pyrope, diopside, and dolomite), dolomite activities are buffered to values near those of Fig. 1b and c.

If carbonatites are to form from melting of eclogite + carbonate + CO_2 , eclogite + CO_2 must melt at temperatures which will allow significant carbonate activities in the resultant liquid. Melting relationships of eclogite + CO_2 carbonate are not known, so that extrapolation from similar systems is necessary. Melting relationships in peridotite + CO_2 are well known^{1,2}, and the solidus in that system shows a sharp inflection to lower temperatures at about 25 kbar, with the minimum at about 1,200 °C and 27 kbar. Above that pressure, the solidus temperature increases monotonically. Peridotite + CO_2 is not directly relevant to eclogite, but the solidi in related fluid-absent systems with sodium (the joins jadeite-forsterite and jadeite-enstatite) are ~400 °C below the solidi in similar systems without sodium (diopside-forsterite and diopside-enstatite)²². Addition of garnet to these systems must reduce the solidi, but the effect will be minor because garnet participates little in early partial melting^{23,24}. By extrapolating these results to vapour-saturated systems, one may expect eclogite + CO_2 to

melt near 800 °C at 25 kbar. Similarly, eclogite + H_2O begins to melt near 700 °C at 25 kbar (ref. 25), and H_2O and CO_2 are found to depress melting similarly at this pressure²⁶. Thus, one may expect eclogite + CO_2 to melt between 700 and 1,000 °C at pressures of 25 kbar. Both extrapolations suggest that eclogite + CO_2 should melt within the field of eclogite + carbonate stability, and thus suggest that carbonated eclogite could reasonably be the source of carbonatite magmas.

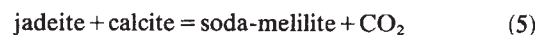
At pressures greater than 25 kbar, partial melts in the peridotite + CO_2 systems are carbonate-rich^{1,2,27}, and it is reasonable to assume that carbonate-rich melts will also result from partial fusion of eclogite + CO_2 at similar pressures. One must then know if partial melts of eclogite + carbonate + CO_2 can have compositions similar to those of natural carbonatites.

As a model for interactions in the natural system, phase relations in the system $\text{NaAlSi}_3\text{O}_4\text{-CaO-SiO}_2\text{-CO}_2$ have been calculated as above (Fig. 2). That system contains most sodium-bearing minerals relevant to eclogite and carbonatite including soda-melilite ($\text{NaCaAlSi}_2\text{O}_7$), which accounts for up to 50 mol% of igneous melilites²⁸. Solid-solid reactions involving jadeite are from Boettcher and Wyllie²⁹ and Holland³⁰, and the free energy of soda-melilite is from the reaction

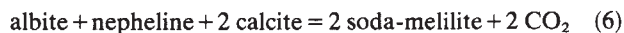


estimated by Yoder²⁸. For clarity, Fig. 2 omits polymorphic transitions and the calcite-quartz-wollastonite- CO_2 equilibrium. Much of Fig. 2 is metastable with respect to melting reactions³¹, but few reactions in the vapour-saturated system are known³².

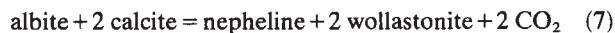
If melting in the system jadeite-calcite- CO_2 is essentially congruent, the resultant melts will contain significant soda-melilite component by



(noted by Kushiro²⁸), although the reaction is metastable with respect to melting relations. If a partial melt in the system jadeite-calcite- CO_2 is emplaced at the Earth's surface (1 bar pressure), it will crystallize to the mineral assemblage nepheline + wollastonite + calcite found in ijolites and some nephelinite lavas. In natural magmas, melilite may be stabilized to 1 bar by akermanite content, and the assemblages melilite ± calcite ± nepheline ± wollastonite will result. These assemblages are characteristic of melilitite lavas and some carbonatites (like that at Oka, Quebec³³). At pressures greater than those defined by the equilibria



and



crystallization of melts in the system jadeite-calcite- CO_2 will produce calcite + albite ± nepheline ± wollastonite rocks. These assemblages may be compared with the carbonatitic alkalic rocks of the Bancroft-Haliburton area, Canada³, feldspar-bearing carbonatites³, or carbonate albitites³⁴. Water content of the mantle fluid, degree of partial melting, and temperature of formation may control the carbonate content of the melt.

The Na_2O and CO_2 content of the melts cannot be estimated because so little is known of the melting relations of carbonated eclogite, and because the original Na_2O and CO_2 content of carbonatite magmas can be affected by fractional crystallization, loss of volatiles to a fenitizing fluid, or later alteration. Carbonatites of the Oka complex (Quebec, Canada) are among the most sodic known, and include rocks with up to 4.5 wt% Na_2O , which is a least upper bound on possible Na content of carbonatite magmas. Partial melts of uncarbonated eclogite^{23,24} contain between 3 and 9 wt% Na_2O , consistent with the sodium content of the Oka rocks.

To investigate the tectonic implications of a genetic connection between carbonatite and mantle eclogite, calculated carbonate/CO₂ activity ratios (Fig. 1) may be compared with calculated geotherms for both continental and oceanic lithosphere³⁵. The continental lithosphere is modelled as an 8 km thick granitic layer ($\rho = 2.7 \text{ g cm}^{-3}$) and a 32 km thick granulitic layer ($\rho = 3.0 \text{ g cm}^{-3}$) overlying mantle eclogite or peridotite ($\rho = 3.4 \text{ g cm}^{-3}$). Oceanic lithosphere is modelled as a 10-km thick gabbroic layer ($\rho = 2.95 \text{ g cm}^{-3}$) overlying eclogite or peridotite. Continental geotherms of 50 and 40 mW m⁻² approximate geothermal gradients in stable cratons and Precambrian shields respectively. The 50 and 40 mW m⁻² oceanic geotherms approximate lithosphere 90 and greater than 120-Myr old respectively³⁶.

If the mantle is generally saturated with a CO₂-rich fluid¹⁹⁻²¹, kyanite eclogite will buffer calcite activities to moderately high values along both plotted continental geotherms and the lower heat flow oceanic geotherm. If kyanite and quartz or coesite are absent, the isopleths of Fig. 1 represent lower limits on carbonate phase activities. Thus, mantle eclogite under areas of low heat flow should be in equilibrium with calcic dolomite or a carbonate-rich fluid or melt, and be a potential source for carbonatites. In fact, carbonatites on continents are restricted to shields and cratons³, and a few are reported from the Cape Verde Islands, presumably on old oceanic crust³⁷, so that calculated-phase equilibria are compatible with the tectonic setting of carbonatites.

Because carbonatite melts are stable with mantle eclogite along normal geotherms corresponding to low surface heat flow, no unusual source of heat is needed to generate carbonatites. In fact, carbonatite melts cannot be stable with eclogite along the high heat-flow geotherms normally ascribed to 'hotspots' or mantle thermal anomalies (Fig. 1). This suggests that carbonatites should not be used as hotspot tracers (as has been done with kimberlites³⁸), but that carbonatite volcanism can be reasonably ascribed to activation of deep crustal fractures³⁹.

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Rapid development of tolerance to the behavioural actions of cholecystokinin

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Cholecystokinin (CCK) acts acutely to inhibit food consumption in fasted rats¹⁻⁴, mice⁵⁻⁷, sheep⁸, pigs⁹, monkeys¹⁰ and humans¹¹⁻¹³. CCK has been proposed as a satiety signal, inducing the behavioural sequence of satiety², or as an aversive internal stimulus, which inhibits food intake by inducing malaise⁴. Reductions in food intake and related exploratory behaviours are initiated by CCK at its peripheral receptor in the gut, which appears to transmit sensory feedback via the vagus nerve to brain regions mediating appetitive behaviours¹⁵⁻¹⁷. The therapeutic potential of CCK as an appetite suppressant in obesity syndromes rests on the demonstration of significant, long-lasting body weight reduction. Chronic CCK administration by repeated injections is problematic, since this peptide is rapidly degraded *in vivo*. We chose the Alzet constant infusion osmotic minipump to investigate possible alterations in body weight and food intake during continuous infusion of CCK. We now report that no change was detected in either body weight or total daily food consumption at any time point during 2 weeks of intraperitoneally (i.p.) infused CCK. The mechanism underlying the lack of chronic CCK effects appears to be a rapid development of behavioural tolerance. Acute challenge doses of CCK which induced satiety-related behaviours in saline-infused rats were ineffective in CCK-infused rats. The behavioural tolerance was apparent within a few hours of minipump implantation. These results provide the first evidence that rapid and reversible tolerance develops to the actions of a gut peptide.

Male Sprague-Dawley rats (200-250 g starting weight), individually housed in a temperature and light-controlled environment, were maintained on an *ad libitum* diet of Purina rat chow and water or condensed milk enriched with multivitamins (Poly-Vi-Sol, Mead Johnson). Alzet osmotic minipumps (Model 2001 1 $\mu\text{l h}^{-1}$, 7 days or Model 2002, 0.5 $\mu\text{l h}^{-1}$, 14 days) were surgically implanted in the abdominal cavity under halothane anaesthesia. Pumps were implanted between hours 4 and 7 of the 12-h lights-on period, when food consumption is normally low, but the gut contains food recently consumed. Fourteen-day pumps were filled with saline or CCK8-sulphate (Bachem) to deliver 0, 0.1 or 1.0 μg per kg body weight per h in experiment 1. Seven-day pumps were filled with saline or CCK8-sulphate to deliver 0, 0.05, or 1.0 μg per kg per h in experiment 2. Filled pumps were pre-warmed to ensure immediate, uniform delivery at the time of implantation. Stability of CCK in the implanted pumps was analysed by HPLC and radioimmunoassay^{18,19} of fluid removed at day 13 from the 14-day pumps. Each animal was weighed daily. Food consumption was recorded in experiment 1 by daily weighing of the rat chow container and in experiment 2 by volumetric measuring of the liquid diet from calibrated bottles which minimized spillage. In experiment 2, satiety-related reductions in exploratory behaviours^{5,20,21} were analysed in 5-min sessions in a video-tracking computer-assisted animal behaviour monitor²², immediately after challenge doses of CCK 5 μg per kg i.p. at 4 h, 1 and 4 days after pump implantation and 1 day after the pump contents were depleted. This monitoring system records the location of the animal as an x-y

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