

Large Igneous Provinces and the Mantle Plume Hypothesis

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Columnar jointing in a postglacial basalt flow at Aldeyarfoss, NE Iceland. PHOTO JOHN MACLENNAN

Mantle plumes are columns of hot, solid material that originate deep in the mantle, probably at the core–mantle boundary. Laboratory and numerical models replicating conditions appropriate to the mantle show that mantle plumes have a regular and predictable shape that allows a number of testable predictions to be made. New mantle plumes are predicted to consist of a large head, 1000 km in diameter, followed by a narrower tail. Initial eruption of basalt from a plume head should be preceded by ~1000 m of domal uplift. High-temperature magmas are expected to dominate the first eruptive products of a new plume and should be concentrated near the centre of the volcanic province. All of these predictions are confirmed by observations.

KEYWORDS: mantle plume, large igneous provinces, uplift, picrite

INTRODUCTION

The plate tectonic hypothesis provides an elegant explanation for Earth's two principal types of basaltic volcanism, mid-ocean ridge and island arc volcanism, both of which occur at plate boundaries. Mid-ocean ridge basalts form new ocean crust along the tensional zones that develop where adjacent plates, with divergent motions, are pulled apart, and island arc magmas form along regions of compression, where plates sink back into the mantle. However, a third significant form of volcanism occurs away from plate boundaries and therefore cannot be explained by plate tectonics. The most volumetrically significant of these are continental flood basalts, giant oceanic plateaus, and aseismic ridges. Continental flood basalts and giant oceanic plateaus, their oceanic equivalent, are massive outpourings of basalt that erupt in 1 to 5 Myr. They cover an equidimensional area typically 2000–2500 km across (White and McKenzie 1989). Collectively they are referred to as Large Igneous Provinces (LIPs). Aseismic ridges are chains of volcanoes that stretch across the sea floor. FIGURE 1 shows the Deccan Traps, a typical flood basalt, and the Chagos–Lacadeive Ridge–Mascarene Plateau, a typical aseismic ridge. Notice that the Deccan Traps are connected by the 200–300 km wide Chagos–Lacadeive Ridge, across the Carlsberg–Central Indian Ridge spreading center, through the Mascarene Plateau to an active volcano at Réunion. The plume hypothesis attributes flood basalts and giant oceanic plateaus to the melting of the large spherical head of a new plume (Richards et al. 1989; i.e. Campbell and Griffiths

1990) and aseismic ridges, like the Chagos–Lacadeive Ridge, to the melting of a plume tail (Wilson 1963; Morgan 1971).

THE MANTLE PLUME HYPOTHESIS

Convection in fluids is driven by buoyancy anomalies that originate in thermal boundary layers. Earth's mantle has two boundary layers. The upper boundary layer is the lithosphere, which cools through its upper surface. It eventually becomes denser than the underlying mantle and sinks back

into it, driving plate tectonics. The lower boundary layer is the contact between the Earth's molten iron–nickel outer core and the mantle. High-pressure experimental studies of the melting point of iron–nickel alloys show that the core is several hundred degrees hotter than the overlying mantle. A temperature difference of this magnitude is expected to produce an unstable boundary layer above the core which, in turn, should produce plumes of hot, solid material that rise through the mantle, driven by their thermal buoyancy. Therefore, from theoretical considerations, mantle plumes are the inevitable consequence of a hot core.

The material in the lower boundary layer will be lighter than the overlying mantle, but before it can rise at a significant rate, it must gather enough buoyancy to overcome the viscosity of the mantle that opposes its rise. As a consequence, new plumes have a large head followed by a relatively narrow tail (FIG. 2). The tail or feeder conduit is comparatively narrow because hot, relatively low-viscosity material following up the existing pathway of the tail requires less buoyancy to rise than the head, which must displace cooler, high-viscosity mantle. As the head rises through the mantle, it grows for two reasons (Griffiths and Campbell 1990). First, material in the high-temperature, low-viscosity tail rises faster than the head and feeds a constant flux of hot mantle into the head of the plume. When this material reaches the stagnation point at the top of the plume, it flows radially with a spiraling motion to give the head its characteristic doughnut shape. Second, the head grows by entrainment. As the plume rises, heat is conducted into the adjacent mantle. In the boundary layers adjacent to the head and tail, the temperature increases and the density decreases, eventually becoming the same as the plume; the boundary layers then begin to rise with the plume. This material becomes part of the plume and is swept into the base of the head by its recirculating motion. The head is therefore a mixture of hot material from the

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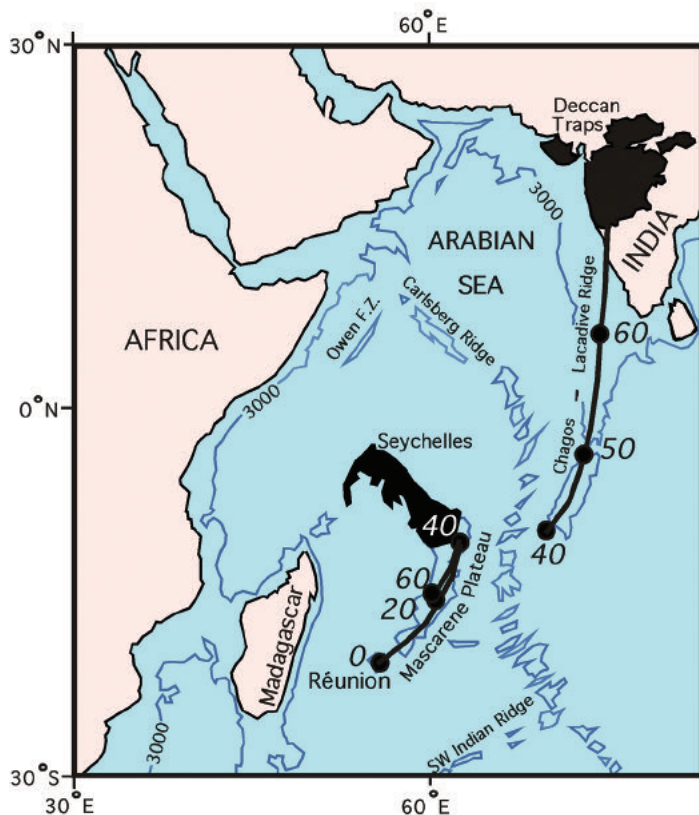


FIGURE 1 Map of the western Indian Ocean showing the distribution of volcanic rocks associated with the Réunion–Deccan plume. The Seychelles were part of the Deccan Traps prior to separation caused by spreading on the Carlsberg–Central Indian Ridge. Note that the Deccan Traps are connected via the 200–300 km wide Chagos–Lacative Ridge, across the Carlsberg Ridge spreading center, to the Mascarene Plateau and eventually to Réunion Island, the current position of the plume. Between 60 and 40 Ma, the Réunion plume was located under the Carlsberg Ridge. It produced a volcanic ridge on both sides of the spreading center, before leaving the ridge and appearing to backtrack on itself towards Réunion, which is its current position (adapted from White and McKenzie 1989).

plume source region and cooler entrained mantle (FIG. 2A). When the plume head reaches the top of its ascent, it flattens to form a disk with a diameter twice that of the head (FIG. 2B). Note that growth of the plume head as it rises through the mantle occurs because mantle in the plume tail rises faster than the plume head, which is a direct consequence of the strong temperature dependence of the mantle's viscosity.

PREDICTIONS AND OBSERVATIONS

The plume hypothesis makes the following testable predictions.

New plumes consist of a large head followed by a small tail

Flood basalts and oceanic plateaus, the oceanic equivalent of flood basalts, are the first eruptive products of a new mantle plume. The volumes of basalt produced during these events are enormous, and Richards et al. (1989) have shown that eruption rates of flood basalts are one to two orders of magnitude higher than those of the associated ocean island chain, which connects them to the current position of the plume. This observation fits well with the plume hypothesis, which attributes the high eruption rates of flood basalts to melting of plume heads and the lower eruption rates of ocean island chains to melting of narrower plume tails.

Flattened plume heads should be 2000 to 2500 km in diameter

The diameter of a plume head (D) depends on the temperature difference between the plume and the adjacent mantle (ΔT , the plume's excess temperature), its buoyancy flux (Q), the kinematic viscosity of the lower mantle (ν), and its height of rise (Z), as described in equation (1)

$$D = Q^{1/5} (\nu/g \alpha \Delta T)^{1/5} K^{2/5} Z^{3/5} \quad (1)$$

where g is gravitational acceleration, α is the coefficient of thermal expansion of the mantle, and K is its thermal conductivity (Griffiths and Campbell 1990). Note that the plume height of rise is raised to the power $3/5$, whereas most other terms are raised to the power $1/5$. Therefore the height of rise of the plume, which in the case of Earth is the depth of the mantle, is the dominant factor influencing the size of a plume head. If ΔT of a plume is assumed to be 300°C and its buoyancy flux to vary between 3×10^3 and $4 \times 10^4 \text{ N s}^{-1}$, the calculated diameter of a plume head originating at the core–mantle boundary is 1000 to 1200 km. The plume head should flatten to produce a disk with a diameter between 2000 and 2400 km when it reaches the top of the mantle. The calculated time that a plume head takes to rise from the core–mantle boundary to the top of the mantle is about 100 Myr.

When a plume head rises beneath continental crust, the associated buoyancy anomaly lifts the lithosphere, placing it under tension. This can lead to runaway extension and to the formation of a new ocean basin (Hill 1991). During the initial stages of rifting, the hot mantle in the underlying plume head is drawn into the spreading centre to produce thickened oceanic crust. If the line along which the continent splits lies close to the centre of the plume head, as was the case when the North Atlantic opened above the Iceland plume head, the length of thickened oceanic crust should be equal to the diameter of the plume head – 2000 to 2400 km. FIGURE 3 shows the zone of thickened oceanic crust along the east Greenland coast and the Rockall–Vøring plateaus, which are associated with the break-up of the North Atlantic. Notice that the zone of thickened oceanic crust on both sides of the North Atlantic is ~2400 km long. When a new oceanic ocean basin opens above a plume head, the first oceanic crust to form is always anomalously thick, typically at least twice as thick as normal oceanic crust. Other examples of plume-related thickened oceanic crust are the margins of South America and Africa above the Paraná plume and India above the Deccan plume (White and McKenzie 1989). Anomalous thickening of oceanic crust does not occur when a new ocean basin forms away from the path of a plume.

Plumes must originate from a hot boundary layer, probably the core–mantle boundary

The obvious way to show that plumes originate from the core–mantle boundary is to use seismic methods to trace plume tails from the top of the upper mantle to their source. However, the small diameter of plume tails (100 to 200 km in the upper mantle, although wider in the lower mantle) makes them difficult to resolve and, as a consequence, attempts to use seismic methods to image plume tails have met with little success. A notable exception is the pioneering use by Montelli et al. (2004) of a new finite-frequency technique to resolve plume tails. They have used this method to trace the Ascension, Azores, Canary, Easter, Samoa, and Tahiti plumes to the core–mantle boundary. A number of other plumes, including the Iceland plume, disappear in the lower mantle, and the Yellowstone plume has no resolvable signature. Montelli et al. (2004) note that it is more difficult to image plumes in the lower mantle than in

the upper mantle, which may explain the discrepancies between the Montelli et al. (2004) images and those expected from plume theory. Nevertheless the Montelli et al. method shows great promise and may eventually allow unambiguous imaging of plume tails in both the upper and lower mantles.

Both heads and tails should erupt high-temperature picrites

The ΔT of mantle plumes can be estimated from the maximum MgO content of their erupted magmas because, for dry melts, there is a linear relationship between the MgO content and magma temperature. As a rough rule, a 4 wt% increase in MgO in the melt equates to a 100°C increase in mantle potential temperature (see glossary). The maximum MgO content of plume-derived picritic (high-Mg basaltic) liquids varies between 18 and 22 wt%, suggesting that the temperature excess for mantle plumes is between 150°C and 250°C. Examples of volcanic provinces that have been attributed to plumes and that include high-MgO picritic melts are Réunion–Deccan, Paraná, North Atlantic Province, Karoo, Emeishan, Galapagos–Caribbean, and Hawai‘i.

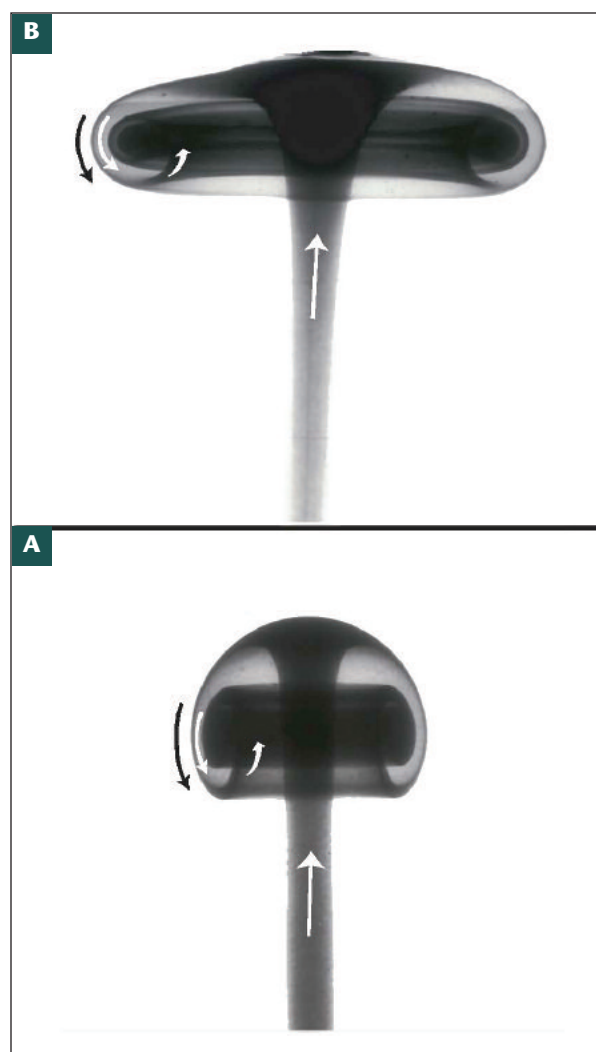


FIGURE 2 Photograph of a laboratory model of a starting thermal plume (A) mid-way through its ascent and (B) after the head flattens at the top of its ascent. The dark fluid represents hot material from the plume source and the lighter fluid is cooler entrained material. White arrows show motion within the plume and black arrows the direction of motion in the boundary layer adjacent to the plume; the boundary layer has been heated by conduction so that its density is approximately the same as that of the plume (after Griffiths and Campbell 1990).

The temperature excess of a plume head is highest at the centre of the head and decreases towards the margin

When new oceanic crust opens up above a plume head, the thickness of the oceanic crust produced is dependent on the temperature of the mantle that is drawn into the new mid-ocean ridge spreading centre. The plume hypothesis predicts that the temperature of a plume head should be highest near the plume axis, where the tail continues to rise through the centre of the head (FIG. 2). At the center of the head, ΔT is expected to be $300 \pm 100^\circ\text{C}$. The ΔT in the remainder of the head, which is a mixture of hot material from the boundary layer source and cooler entrained lower mantle, varies with the plume buoyancy flux but must be less than at the centre. For typical plume buoyancy fluxes, the average ΔT is $\sim 100^\circ\text{C}$. Hopper et al. (2003) have used seismic reflection and refraction data, obtained from four traverses located between the plume head and its margin, to determine the thickness of the first oceanic crust to form when the North Atlantic opened above the Iceland plume. They obtained thicknesses of 33 and 30 km for traverses T-I and T-II, close to the plume axis, and 18 and 17 km for traverses T-III and T-IV, closer to the margin of the head (see FIG. 3). The ΔT required to produce these crustal thicknesses, based on the work of McKenzie and Bickle (1988), is 350°C and 100°C at the centre and margin of the head, respectively, which is consistent with the plume hypothesis.

Picrites should erupt early during flood volcanism and be most abundant near the centre of the plume head and less abundant towards the margin

The hottest material in the head is the mantle from the plume source (the dark colored fluid in FIG. 2), which is $300 \pm 100^\circ\text{C}$ hotter than the entrained mantle. Although this temperature difference decreases with time, as adjacent layers exchange heat, the high temperatures of the hot layers will persist for millions to tens of millions of years, depending on their thickness, which in turn depends on their distance from the plume axis (see FIG. 2). When a plume head melts to form a flood basalt province, only the top of the head ascends to a level in the mantle where the pressure is low enough to allow melting. Note in FIG. 2 that the hot layer at the top of the plume thickens towards the centre, where it grades into the tail. Provided the first magmas for a new plume do not pond and fractionate in crustal magma chambers, picrites should dominate the early melting products of plume heads and become less abundant as the cooler, second layer ascends to a level where it can begin to melt. Picrites should also be most abundant towards the hot centre of the plume head and become less abundant towards the margins. The predicted early picrites have been documented for the Paraná–Etendeka, Deccan, Emeishan, North Atlantic, and Karoo flood basalts. Examples of picrites that are abundant at the centre of a flood basalt province and less abundant towards the margins are the Letaba picrites, as seen along the Lobombo monocline of the Karoo, and those in the deeply dissected valleys of the Emeishan flood basalt province in China.

Flood volcanism should be preceded by domal uplift of 500 to 1000 m at the center of the dome

The arrival of the hot plume head in the upper mantle will produce domal uplift at the surface, the magnitude of which depends on its average temperature. The area of maximum uplift is predicted to have a radius of ca. 200 km and to be surrounded by a zone, with a radius of ca. 400 km, in which uplift is still significant (FIG. 4).

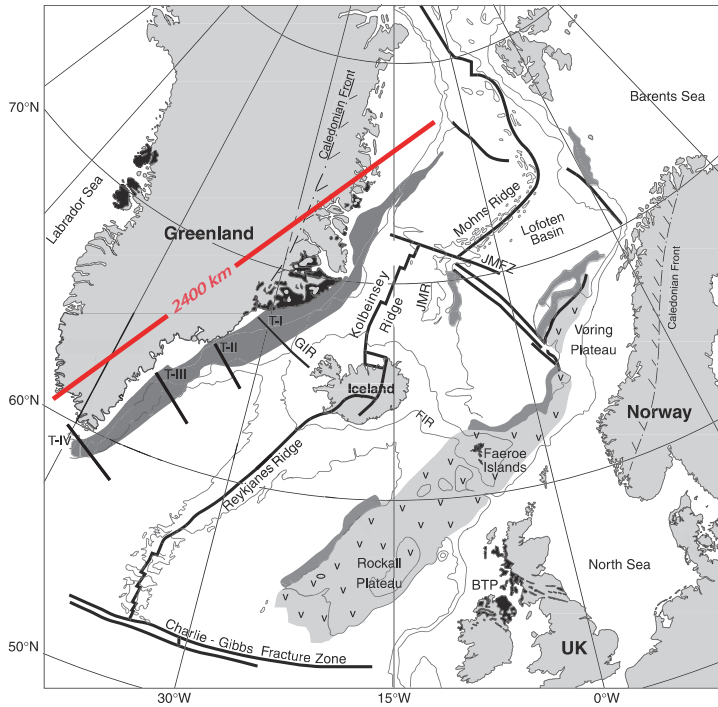


FIGURE 3 Geological map of the North Atlantic region after Hopper et al. (2003). Areas shaded black correspond to onshore basalts, which erupted from the plume head prior to the opening of the North Atlantic. The dark grey areas locate offshore seaward-dipping reflectors, indicating areas of thickened oceanic crust. Light grey area with v's between the Rockall Plateau and the Vøring Plateau shows the areal extent of basalts associated with the initial opening of the ocean. T-I to T-IV are seismic transects along the East Greenland coast (from Hopper et al. 2003). BTP: British Tertiary Province; FIR: Faeroes-Iceland Ridge; GIR, Greenland-Iceland Ridge; JMR: Jan Mayen Fracture Zone.

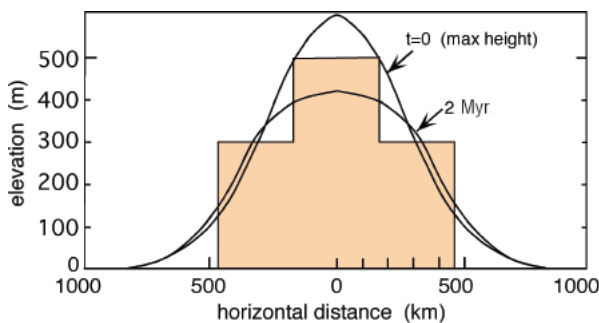


FIGURE 4 Uplift above a plume head, as predicted by Griffiths and Campbell (1991), compared with the uplift observed at the centre of the Emeishan flood basalt (shown in pink) by He et al. (2003). The timing and uplift shape predicted by Farnetani and Richards (1994) is similar, but they predict more uplift because they model a plume head with a higher excess temperature: 350°C as opposed to 100°C. Predicted profiles are given for maximum uplift ($t = 0$), when the top of the plume is at a depth of ~250 km, and 2 Myr later ($t = 2$ Myr), when flattening of the head is essentially complete. The uplift for the Emeishan flood basalt province is the minimum average value for the inner and intermediate zones as determined from the depth of erosion of the underlying carbonate rocks.

The best-documented example of domal uplift occurs in association with the Emeishan flood basalt province in China (He et al. 2003). Here the carbonate beds of the underlying Maokou Formation have been systematically thinned by erosion towards the centre of the flood basalt province. Isopachs of the Maokou Formation show that uplift was broadly dome shaped as expected. He et al. recognize three zones: an inner zone with a radius of ca. 200 km, where uplift is estimated to be at least 500 m and could pos-

sibly exceed 1000 m; an intermediate zone with a radius of 425 km, in which the average uplift is ca. 300 m; and an outer zone with a radius 800 km, in which uplift is minimal. The magnitude and shape of the uplift agree remarkably well with that predicted by Griffiths and Campbell (1991) (Fig. 4) and Farnetani and Richards (1994). Other examples of uplift prior to volcanism include the Wrangellia province of northwest Canada and southwest Alaska, the Natkusiak province in northwest Canada, the Deccan flood basalt of India, and the Ethiopian flood basalt. Uplift started 3 to 5 Myr before the Emeishan and Wrangellia volcanism.

THERMOCHEMICAL PLUMES

An obvious weakness of the thermal plume hypothesis is that it fails to explain a number of minor volcanic chains that stretch across the ocean basin and cannot be linked to LIPs. These appear to be the product of plume tails without heads. It has been suggested that the plumes responsible for these volcanic chains may originate in the mid-mantle. Theoretically the mantle could divide into two convecting layers separated by a boundary layer at the 670 km seismic discontinuity; this boundary layer is a potential source of plumes, which would rise a small distance and therefore have small heads. However, seismic studies and numerical models showing that slabs and plumes, respectively, can pass through the 670 km discontinuity make this interpretation unlikely. A recent computer model of thermocompositional plumes (Farnetani and Samuel 2005) provides a possible solution. They have shown that if the head of a weak plume contains a high proportion of a dense mantle component (e.g. subducted former basaltic crust), the ascending head can stall at the 670 km discontinuity and separate into low- and high-density components. Only the light component penetrates the discontinuity and gives rise to a weak plume with a smaller temperature excess than the strong plumes associated with LIPs. The suggestion that plumes may have an above-average mantle concentration of dense, subducted former oceanic crust is consistent with the usual interpretation of D'' – a heterogeneous, seismically fast layer of variable thickness (300 ± 300 km) at the base of the mantle – as a zone where subducted slabs have concentrated. It is also consistent with the isotopic (both stable and radiogenic) and trace element characteristics of many plume-derived magmas, which also require a higher than average mantle concentration of former oceanic crust.

CONCLUSION

The mantle plume hypothesis provides a simple explanation for all of the essential features of classic LIPs, such as the Deccan Traps and the North Atlantic Igneous Province, and all of its predictions have been confirmed by observation. Five days of intense scrutiny during The Great Plume Debate (Fort William, 28 Aug–1 Sept 2005) failed to land a telling blow on the mantle plume chin, and no credible alternative emerged to explain the principal features of LIPs, especially their high eruptive volumes and high-temperature magmas. However this does not mean that all intraplate volcanoes and volcanic chains are due to plumes. Each case must be considered on its own merits.

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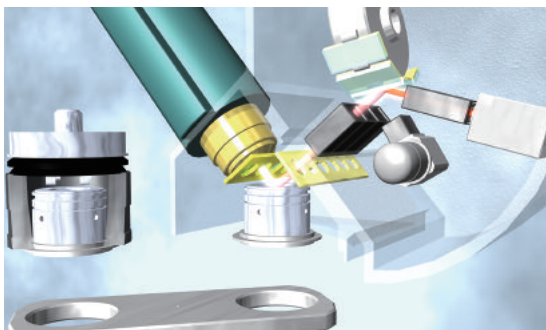
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