Thermally mantled gneiss domes: the case for convective heat flow in more or less solid orogenic basement

by E. den Tex

Summary

The characteristics of mantled gneiss domes and thermal dome structures are reviewed. Two examples of thermally mantled gneiss domes, where the two features are combined in a single dome, are discussed in more detail. They are the Agout dome in central France and the Lepontine gneiss dome of the Swiss Alps. Various models proposed to explain the mechanism of formation of thermally mantled gneiss domes are reviewed and their applicability to the domes under scrutiny is investigated.

It is concluded that the conductive/convective model of Talbot offers a better fit to their characteristic features than the radioactive/conductive and magmatic/conductive models. The relation between thickness of source layer, metamorphic facies series, and the nature and extent of thermal basement convection is briefly explored.

1 Mantled gneiss domes

Mantled gneiss domes are dome-shaped structures consisting of a generally conformable gneissose core enveloped by a predominantly concordant mantle of supracrustal rocks such as (meta-)sediments and (meta-)volcanics. The concept of the mantled gneiss dome has come into prominence mainly through the work of Eskola (1949). He argued that mantled gneiss domes are characteristically poly-orogenic features because the boundary between gneissose core and supracrustal mantle is always marked by a conglomeratic, quartzitic or dolomitic horizon. According to Eskola the structural concordance between these originally discordant units of the mantled gneiss dome is due to migmatization and granitization, with concomitant remobilization and squeezing of the basement up into the overlying supracrustals during a second orogenic cycle. He admitted that granitic rocks intrusive into the supracrustal mantle are present in some mantled gneiss domes, but he considered them to be post- or late-second cycle features. Indeed mantled gneiss domes appear to be restricted to ensialic orogens, in which pre-existing continental crust was recycled. Apart from the Karelidic belt in the Baltic shield of eastern Finland and western Russia (Eskola, 1949), they have been reported from the Grenville/Taconic/Acadian orogen of the Appalachian belt that runs from Maryland

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through New Jersey, Connecticut, Massachusetts, New Hampshire and Vermont into Maine (Chapman, 1942; Lyons, 1955; Tilton et al., 1958; Hopson, 1964; Wetherill et al., 1968; Thompson et al., 1968; Naylor, 1969), from the Albion Range in the Basin and Range Province, southern Idaho (Armstrong, 1968), from the Panamint Range near Death Valley in SE California (Lanphere et al., 1964), from the Variscan Pyrenees in France and Spain (Fonteilles and Guitard, 1968a, b; Zwart, 1968) from the Karagwe-Ankolean belt in southern Uganda and the Umkondo-Zambesi belt in N.E. Rhodesia (Nicholson, 1965; Talbot, 1971), and from the Shuswap terrain of SE British Columbia (Hyndman, 1968; Reesor, 1970). Most of the recently published data on mantled gneiss domes are in agreement with Eskola's genetic model. Core and mantle are generally present in a basement-to-cover relationship which is more or less obscured by transposition of the original unconformity. Isotopic age determinations have confirmed the presence of polycyclic basement in the core of domes mantled by monometamorphic cover rocks, and intruded by post-metamorphic granitic rocks. There are, however, some significant exceptions. For instance, in the eastern Appalachians the Oliverian domes contain stratified and unstratified core rocks ranging from granite to quartzdiorite in composition (the so-called Oliverian magma series) which were shown to be formed and emplaced pene-contemporaneously with the immediately overlying Ammonoosuc volcanics of Middle Ordovician age (Naylor, 1969). The mantle rocks proper, the lowermost components of which are conglomerates and quartzites of the Silurian Clough and Fitch formations, were metamorphosed only in Middle Devonian times. Lanphere et al. (1964) have also presented evidence against remobilization of the basement core in the World Beater Complex of SE California. There a granite has intruded the basement core but was truncated by the mantle rocks. Naylor (1969) has found similar relationships in the Mascoma dome of Western New Hampshire. However, in the Western Appalachians and in Maryland, one finds mantled gneiss domes that contain Precambrian core rocks, such as the Baltimore gneiss in the Maryland domes. These domes are concordantly mantled by the late-Precambrian Glenarm series which yields metamorphic mineral ages between 350 and 287 my (Tilton et al., 1958; Hopson, 1964; Wetherill et al., 1968). Thus, before rejecting Eskola's model, it seems worthwile trying to adapt it to allow for more than one subsequent orogeny involving pre-existing crystalline basement. This condition obviously prevailed in the polycyclic Grenville/Taconic/Acadian belt of the Appalachians.

2 Relations between gneiss domes and thermal structures

Regional thermal domes are known from a diversity of metamorphic terrains. They occur in the low- and intermediate- but never in the high-pressure facies series (Miyashiro, 1961). Further, they occasionally appear to form more or less regular linear or grid-patterns centred on the axial zone of an orogenic belt. Representatives of such patterns are the ten domes, spaced at roughly 25 km intervals, in the Pyrenees (Zwart, 1968; Fonteilles and Guitard, 1968), and the dozen or so domes roughly 25 km apart in the Northern Appalachians (Thompson et al., 1968). The structure of these thermal domes is delineated by the outcrop patterns of isograde surfaces which are determined by the incoming or outgoing of metamorphic index minerals, or by the presence of reacting mineral assemblages (Winkler, 1974). With the highest grade of metamorphism located in their centre, the thermal domes may be interpreted as local bulges of the normally subhorizontal geo-isotherms.

Frequently the thermal dome structure coincides with a mantled gneiss dome, i.e. the high-grade core of the thermal dome consists of recycled gneiss and granite, and the lower grade mantle of monocyclic supracrustal rocks. A clear example of such a thermally mantled gneiss dome is the 'dome de l'Agout' in the 'Montagne Noire' at the SW perimeter of the French Central Massif (Roques, 1941; Schuiling, 1960; Arthaud, 1970). This dome measures roughly 60 by 25 km, its long diameter being parallel to the late-Variscan structural trend (fig. 1). Its emplacement postdates the formation of early Variscan nappe structures of Peninnic style (Arthaud, 1970). The core of the dome consists of peripheral ortho- and paragneisses, the former yielding a Rb/Sr whole rock isochron at 530 my, and a central migmatite terrain which generally produces Rb/Sr whole rock isochrons at 419–475 my, but which is locally re-homogenized at 280–320 my (Roques et al., 1971; Hamet and Allègre, 1972). The mantling metasediments range from Upper Brioverian to Lower Carboniferous with significant disconformities in the Upper Cambrian (Sardic phase) and at the Ordovician/Silurian junction (Taconic phase). Several metamorphic isograds have been recognized in the area of the dome (Schuiling, 1960; Bogdanoff, 1973). These are roughly parallel to the core-mantle boundary, which is generally concordant. The inward sequence of isograds is: biotite, garnet, andalusite, staurolite, sillimanite I (+muscovite), sillimanite II (—muscovite) and cordierite, thus defining a low-pressure intermediate or Michigan-Idaho type of facies series (Miyashiro, 1961; Hietanen, 1967). According to Schuiling (1960), who only studied the eastern half of the dome, the outer four metamorphic zones cover a depth of 500 m in the 'micaschistes inférieures', and the two sillimanite zones a similar thickness in the basement rocks. The sillimanite I-isograd coincides almost exactly with the boundary between micaschists and ortho/paragneisses, while the sillimanite II-isograd approximately marks the beginning of anatexis in these gneisses, and the cordierite isograd more or less delineates a central body of anatexitic gneisses and granites (fig. 1). From the incoming of biotite to the beginning of melting in granitic rocks a temperature increase of 250 $\pm$ 50°C, depending on P_{H_2O}, is achieved within a rock column of roughly 1 kilometre suggesting a local thermal gradient in the range of 200–300°C/km (cf. Schuiling, 1960). Such a gradient is abnormally steep for regional metamorphism, and is probably compensated by a much more gentle gradient in the overburden of the Agout dome. This can be verified, since the muscovite-out isograd (Sil II) roughly coincides with the beginning of melting in the ortho- and paragneisses of the outer core mass. If the system was water-saturated, the boundary curves of the reactions:

$$\mathrm{Ms} + \mathrm{Q} \rightleftharpoons \mathrm{Kfs} + \mathrm{Sil} + \mathrm{W}, \text{ and:}$$

$$\mathrm{Ab} + \mathrm{Or} + \mathrm{Q} + \mathrm{W} \leftrightharpoons \text{aplitic melt}$$

must have intersected where these isograds coincide indicating that the temperature was in the range 650–675°C and P_{H_2O} (= P_l) about 3–4 kb, corresponding with depths between 11 and 15 km (Winkler, 1974). Thus the overall geothermal gradient must have been in the range 43 to 60°C/km steepening to 250°C in the lowermost kilometre of the crustal column exposed in the Agout dome. This would leave a gradient of 32–45°C/km for the uppermost 10–14 km, which is significantly steeper than the average conductive geotherm in the continental crust (fig. 2). However, if the system was water-deficient, the

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Sil II (Ms-out)-isograd might mark the beginning of incongruent melting by the forward reaction:

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while the cordierite-isograd might indicate the forward reaction:

$$\text{Bi} + \text{Sil} + \text{Q} \rightleftharpoons \text{Kfsp-melt} \pm \text{Alm-ga} + \text{Cord} + \text{W}$$

(Brown and Fyfe, 1970; Winkler, 1974). Since cordierite co-exists with garnet in the Agout-anatexites, P_l should be higher, in the range of 4–7 kb (corresponding with depths between 15 and 26 km), while the temperature of cordierite appearance would increase to approximately 750°C, yielding an overall geothermal gradient of 24–31°C/km with the lowermost kilometre at 300°C and the uppermost 14 or 19 km at 24–31°C/km. The occasional occurrence of kyanite in the staurolite zone (Collomb et al., 1962) suggests that the depth exceeded 15 km at 450°C, while the general presence of andalusite in the dome envelope indicates that it was metamorphosed at a depth of not more than 19 km. Moreover, the coincidence of the Sil II- and Cord-isograds with zone boundaries of differential anatexis may reflect the stepwise nature of incongruent melting reactions in the water-deficient system of the core gneisses.

In either case it remains to be resolved what caused the sudden steepening of the geothermal gradient over the Agout dome. Was it conductive or convective heat flow or both, or was it heat flow derived from a body of magma intruded into the dome but largely hidden below the present surface? We will attempt to answer these questions at a later stage.

Another thermal dome structure with a gneissic core, petrologically even more thoroughly investigated than the Agout dome, is the Lepontine gneiss region of the Swiss Pennine Alps (figs. 3 and 4). It measures roughly 100 × 50 km, and is situated in the axial culmination of the Pennine nappes. It is enveloped by Mesozoic cover rocks and by more or less telescoped isograds of the late-Alpine or Lepontine phase of metamorphism, representing an intermediate-pressure or Barrovian facies series from the 'stilpnomelane-out' via the 'staurolite-in' and 'kyanite-in' isograds in pelitic rocks of the mantle, to the sillimanite isograd in gneisses of the core mass (Niggli, 1970; Wenk, 1970; Frey et al., 1974). Possibly due to the lack of sufficiently detailed structural knowledge, the Lepontine gneiss region has never explicitly been discussed in terms of the concept of mantled gneiss domes. Nevertheless, there is some consensus of opinion among students of the Lepontine Alps concerning the following points:

a. The Lepontine thermal dome structure postdates the Penninic nappe structures and the Eo-Alpine high-pressure metamorphism (Niggli, 1970; Wenk, 1970; Thakur, 1973; Jäger, 1973).

b. The core gneisses must have been remobilized, since they are generally concordant and conformable with the mantling Mesozoic cover rocks, even though they have predominantly remained closed systems as far as whole rock Sr-isotope exchange is concerned. In fact they predominantly yield Rb/Sr whole rock isochrons of Variscan or rejuvenated Variscan age (Hunziker, 1970; Jäger, 1973).

c. Alpine pegmatites, aplites, migmatites and granites are comparatively scarce and increasingly so in the order mentioned. They are largely confined to areas in the core gneisses enclosed by the sillimanite isograd or exhibiting abundant development of cordierite and wollastonite (fig. 3). These areas

appear to grossly coincide with five probable subdome structures of a lower order: the Toce- and Ticino-culminations (Preiswerk, 1921), the periclinal structures with dominant downdip-lineation in the upper Verzasca-, Maggia- and Peccia valleys (Günthert, 1954; Thakur, 1973), the broad-crested antiform with migmatite subdomes between the Misox and Mera valleys (Wenk, 1956; Blattner, 1965), and the Bergell dome (H. R. Wenk, 1973). They seem to be regularly spaced, some 25 km apart, and to be chiefly aligned along the steeply dipping 'root-zone' of the Pennine nappes (figs. 3 and 4; cf. Den Tex, 1974).
d. The average Lepontine geothermal gradient was probably between 30 and 52°C/km (Clark and Jäger, 1969; Wenk, 1970) steepening towards the gneissose core and especially towards the late-Lepontine Bergell 'granite'.
e. Differential post-metamorphic uplift of the Lepontine gneiss region, as evidenced by the cooling ages of micas (Jäger, 1973; Frey et al., 1974) has probably disturbed the original disposition of the isograds.

The latter condition precludes the derivation of the Lepontine geotherm from the true distances between isograds of approximately known temperature. One should therefore discredit an attempt (Den Tex, 1974) along these lines, which suggested a geothermal gradient steepening from 5°C/km in the envelope to 20°C/km in the core. A better approximation of the geothermal gradient is obtained by trying to establish restricted fields or points in the PT-graph defined by intersecting boundary curves of mineral reactions. In the Lepontine dome the staurolite- and kyanite-isograds virtually coincide. This means that temperatures of about 550°C were reached at a depth of at least 18 or 23 km, depending on the preferred position of the andalusite/kyanite boundary curve (Winkler, 1974). On the other hand cordierite has been reported in apparent co-stability with kyanite in a subsolidus assemblage from the Verzasca Valley (Wenk, 1968), while sillimanite, and not kyanite, is the stable aluminosilicate in the presence of pegmaplitic melt. These data narrow the possible depth range for the climax of Lepontine metamorphism to the interval between 19 and 25 km at approximately 650°C. Thus, it may be concluded that the overall Lepontine geothermal gradient was somewhere between 26 and 34°C/km, having a low rate of 24 to 30°C/km down to the level at which staurolite and kyanite were first formed, but from thereon increasing to between 50 and 100°C/km until the level of partial granitic melting was reached.

Various models have been invoked to explain the dynamic and thermal history of the Lepontine gneiss region (e.g. Jäger, 1962; Niggli, 1970; Wenk, 1970; Winkler, 1970; Den Tex, 1974). They vary from purely conductive flow of magmatic or radioactive heat to interrelated conductive and convective heat flow from various sources in the crust or in the upper mantle. The specific applicability of these models to the thermal domes in the Lepontine and Agout regions will be discussed in the following paragraphs.

3 The radioactive/conductive model

One model attempting to explain thermally mantled gneiss domes uses radioactive decay as the sole heat source and conductivity as the exclusive transfer mechanism of heat to the surface. It is generally admitted that radioactive decay is the major heat source of the earth, and that it is strongly concentrated, but rather evenly distributed within the continental crust. Thus the radioactivity of a column of granitic crust, 30 km thick, can account for more than the measured average heat flow of the earth. The effective flow of heat by conduction (Q) is a linear function of the temperature difference

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The thermal conductivity of crustal rocks is extremely low. Also, at constant temperature, it does not vary beyond one order of magnitude (Clark, 1966).

On these grounds it has been assumed that the continental crust is an infinite plate of granitic material of uniform radioactivity and heat conductivity. If no heat is transferred from the mantle to the crust, the steady-state temperature distribution should have a gradient that progressively declines with depth (Verhoogen et al., 1970). This is the type of geothermal gradient responsible for high-pressure or burial metamorphism (cf. Precambrian shield geotherm, figs. 2 and 5). The fact that the lower continental crust is relatively depleted in the radioactive parent nuclides (U, Th, K, Rb) with respect to the upper continental crust should emphasize this tendency.

According to this model the steady-state geo-isotherms are arranged parallel to the crustal plate boundaries but their spacing increases with depth (Verhoogen et al., 1970; Turner, 1968). A column, 30 km deep, consisting of granite of average radioactive heat productivity and conductivity should under these conditions possess an average geothermal gradient of 40°C/km.

This model does not fit some of the principal characteristics of thermally mantled gneiss domes, such as the Agout dome and the Lepontine gneiss dome. These structures display essentially thermal isograds, of pronounced dome shape, becoming more closely spaced towards the centre of the dome, and at originally deeper levels. They imply geothermal gradients that steepen downwards from 20 or 30°C/km to well over 200°C/km. It is true, that gneiss domes may be due to subsequent isostatic uplift of a locally thickened granitic crust (Niggli, 1970; Jäger, 1973), but it requires too much special pleading to explain the regular telescoping of the isograds by the same mechanism.

Fonteilles and Guitard (1968a, b) modified the conductive model to match the telescoped dome shape of isograds by a so-called basement effect ('effet de socle'). The constraints of their model are: a crustal plate of uniform heat flow at the base, but of variable composition, thickness and heat conductivity. Decreasing thermal conductivity with depth, where the temperatures are higher (but still below the radiation threshold), causes the geothermal gradient to steepen, i.e. to increase its rate at greater depth. Loss of heat suffered by the systems involved in dehydration and partial melting reactions has a similar effect on the disposition and spacing of isograds and on the rate of the geothermal gradient. Except where anatexis is involved, a pre-existing 'dry' basement should suffer much less from these dehydration effects than the non-metamorphic 'wet' sediments and volcanics of the envelope. Rejecting the steady-state model for most metamorphic isograds, Fonteilles and Guitard attribute the telescoping of isograds especially above the top of a basement dome to the more rapid ascent of the geo-isotherms in sectors where the thickness ratio of basement to cover is high.

Undoubtedly this more sophisticated model, rigorously treated by Fonteilles and Guitard, is capable of explaining many features of the thermal structure of mantled gneiss domes. However, it fails to comply with some basic properties of the domes in question. First and foremost is the fact that the model requires the domes to be present – at least in embryonic form – before the basement can effectively create telescoped thermal dome structures around them. It requires very special circumstances to have regular networks of such domes available in the axial zone of orogenic belts such as the Pyrenees and the Appalachians. Secondly, it fails to explain the general concordance between basement-core and mantling supracrustals, irrespective of whether or not the core-envelope boundary rocks were subject to anatexis. Finally it is doubtful if the time-dependent nature of the isograds could be fully responsible for the extent to which they are telescoped, and for the steepening of the geothermal gradient (from 30 to 300°C/km), as observed for instance in the Agout dome. Even when it is admitted that the total amount of heat absorbed by the dehydration of biotite and muscovite and by the melting of granitoid rocks is equal to that needed to raise the rock temperature by 200°C, this would only steepen the average geothermal gradient of the Precambrian shield from 15 to 25°C/km.

4 Magmatic/conductive models

The radioactive crustal plate, acting as heat source in the previous models may be replaced by a body of magma intruded into a sequence of gneisses and supracrustal rocks. Again, the sole transfer mechanism of heat is assumed to be conduction through solid rocks. Jaeger (1964, 1968) has developed a number of such models, for instance for a tabular sheet and for a cylindrical pipe of given thickness or diameter, and of infinite length. These consist of magma injected instantaneously at liquidus temperatures into rocks of uniform thermal properties, which are subject to a given regional temperature gradient. The tabular sheet model may be applicable to linear arrays of thermal domes, such as the Pyrenean and Appalachian gneiss domes, while the cylindrical pipe model is more appropriate to 'single' dome structures such as the Lepontine and Agout domes. In both cases the geo-isotherms develop a more or less elongate dome-shaped configuration over the magma-body, and they collapse gradually as cooling proceeds. Taking into account the effect of the latent heat of magmatic crystallization and making reasonable assumptions for the parameters involved, Jaeger showed that the temperature in the adjoining country rock is a function of the ratio between distance from the contact (X) and diameter of the magmatic pipe (D), while the time (in years) during which the country rock is held at the maximum temperature is approximately $0.01 \times D^2$ (D in metres; Winkler, 1974).

Jäger (1962) rejected the tabular sheet model of granite held at the minimum magmatic temperature (T_m) of 700–500°C for the cooling phase of the Lepontine gneiss dome, assuming 5 km for D and deriving a time interval of 1 my for the centre of the intrusion to cool down to 300°C – the approximate blocking temperature of biotite for Sr-isotope exchange. The difference of 5 my found by Jäger for the cooling ages of biotites is too large to be accounted for by a local magmatic heat source, while the areal distribution of biotite cooling ages shows no coherence with the loci of maximum Lepontine heat flow.

If we apply the cylindrical pipe model to the Agout dome, the following constraints can be assumed: the magma in the pipe has a maximum temperature (T_m) of 800°C, a maximum diameter (D) of 20 km, and a maximum depth of overburden (O) of 26 km, while the metamorphic aureole (from biotite-to cordierite-isograd) is at the most 1 km thick (cf. pp. 64). Under a regional geothermal gradient of 15°C/km the temperature prevailing at the igneous contact (presumably located at the cordierite isograd) should then be 64%

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of the magmatic temperature (T_m) plus the local country rock temperature (T_c) of 390°C, i.e. 900°C; and at 1 km from the presumed igneous contact (near the biotite isograd): 55% of T_m plus T_c (375°C), i.e. 815°C (Winkler, 1974). The latter temperature, especially when maintained for 4 my, is much too high for the biotite isograd which should be situated roughly at 450°C. Calculations on the basis of different constraints such as: $T_m = 650$°C, D = 10 km, O = 11 km, and $T_c = 150$°C (with a geothermal gradient of 15°C/km), still give a temperature at the biotite isograd which is almost 60°C too high.

The conclusion is that pipe-shaped bodies of granitic magma would have to be small and virtually solid, and to be emplaced at very high crustal levels, if they are to be deemed responsible for the disposition and spacing of the metamorphic isograds around the Agout dome. Also, as pointed out by Turner (1968), the isograds around a magmatic pipe should reverse their dip direction and curve back towards the conduit at depths in excess of 10–15 km. No indications of such reversals are available around the Lepontine and Agout domes.

5 The convective/conductive model

The models considered so far neglect the role of thermal convection in crustal heat flow. Jaeger (1964) discussed the influence of convection on the conductive heat flow from a cooling magmatic intrusion, but he decided that it is negligible for distances more than half the diameter of the igneous body constituting the heat source. However, if the presence of pre-existing nappes of Penninic style leads to the assumption that the orogenic continental crust is a layer of viscous Newtonian 'fluid', heated from below to about half its melting point, then a super-adiabatic temperature gradient should cause gravitational instability of the lower crustal layer, leading either to a stationary, cellular and laminar convective motion, or to an irregular, oscillatory and turbulent one, depending on the extent to which its critical Rayleigh number (R_c) is exceeded. This dimensionless number is defined by:

$$R = (g\alpha\beta d^4)/(\varkappa v)$$

in which g denotes the acceleration due to gravity, α the thermal expansion, β = the adverse temperature gradient at the onset of convection, d the thickness of the active layer, $\varkappa$ the thermal diffusivity, and v the kinematic viscosity. The parameters β and $\varkappa$ account for the effect of heat conduction. When the layer is heated from below, as in the case of a mantle diapir rising from a subducted oceanic plate, the convection cells tend to be hexagonal in cross-section, whereas they become horizontally elongate when the heat source is internal, as in the case of radioactive decay. In the first instance there is a simple linear relation between the thickness of the convecting layer and the spacing of the cells. Assuming rigid boundary conditions, both above and below the layer, the thickness $d \simeq \lambda/2$, where λ is the spacing of cells or in this case the wavelength between domes. In that event the critical Rayleigh number (R_c) is 1,700 (Knopoff, 1964).

In a thought-provoking paper Talbot (1971) reported four orders of sub-domes in a mantled gneiss dome in Fungwi Reserve, Rhodesia. He adjusted the Rayleigh model of Bénard convection (Chandrasekhar, 1961) to suit his case. The general lack of structural disturbance, melting and homogenization phenomena in all but the fourth-order domes led Talbot to assume that only marginal instability occurred in the domes of lower order, giving rise to what he called repetitive, non-cyclic convection. Talbot's concept implies an intimate interplay of thermal conduction and convection. When the Rayleigh number just exceeds the critical value by conduction only, a horizontally foliated gneiss should bulge upwards in the centre of a convection cell. Owing to the anisotropic thermal diffusivity of lineated and foliated rocks (Clark, 1966; Wenk and Wenk, 1969), the steeply inclined margins with downdip-lineations characteristic of a gneiss dome should permit a conductive heat flow, two or more times as large as the horizontally foliated top, bringing convection to a halt in the margins, and storing heat in the top for a subsequent convective half-turn on a smaller scale. Ultimately, as in the fourth-order domes of Fungwi Reserve, this process may lead to cyclic, turbulent convection causing structural disturbance, homogenization and melting of the foliated gneiss in its upper levels.

Can the Talbot model be applied to the thermally mantled gneiss domes under discussion? For the Lepontine thermal gneiss dome the numerical values for the parameters of the Rayleigh equation are rather precisely obtained: $g = 10^3$ cm s^{-2}; $\alpha = 5 \times 10^{-5}$ °C^{-1}; β (before the onset of convection to be equated with the geothermal gradient outside the staurolite + kyanite isograd) $= 2.7 \times 10^{-4}$ °C cm^{-1} (fig. 5); d (half the wavelength between subdomes) $= 12 \times 10^5$ cm; $\varkappa = 0.01$ cm^2 s^{-1}; v (for granitic rocks at not more than half their melting temperature) $\approx 10^{17}$ stokes. Substituting these values in the Rayleigh equation gives a value of 28,350 for R.

The critical value of 1,700 is thus exceeded, but only by one order of magnitude. It appears then, that marginal instability of the layered system of viscous fluids may have been achieved sufficiently to cause non-cyclic laminar convection only, but that does not preclude cyclic turbulent convection in some of the Lepontine subdomes. Indeed, in the top of the Bergell subdome, where the solidus temperature of felsic gneisses (~ 700°C) was clearly exceeded, the value of $\varkappa$ must have been considerably lower than 10^{17}, perhaps between 10^9 to 10^6 stokes (Clark, 1966; Shaw, 1965). In this instance d cannot have been much less than 10^5 cm. Here, R_c may have been exceeded by 5 to 8 orders of magnitude, permitting turbulent convection of the cyclic type, whereby the felsic gneisses were structurally disturbed and thoroughly homogenized to yield the Gruf migmatites and, ultimately, the overlying granodiorite and tonalite of the Bergell Alps. H. R. Wenk (1973) has produced a detailed structural picture of the Bergell Alps. His sections suggest an asymmetrical dome with a diapiric 'hat' of granitic rocks in the NW and a pinched rim-syncline in the SE part. But the overall concordance of the generally gneissose Bergell 'granite' with the migmatites and gneisses of the underlying Tambo nappe led Wenk to date its emplacement as pre-nappe. This deduction is unwarranted.

Apart from β and d, the same numerical values can be considered appropriate for the parameters of the gneiss layer in the Agout dome. The average geothermal gradient before the onset of convection (β) may be estimated at 31°C/km or 31×10^{-6} °C cm^{-1} (fig. 2). The thickness of the convecting layer (d) may be derived from the distance between subdomes, as in the Lepontine case, since the eastern Agout dome consists of two peri-antiformal lobes (Espinouse and Caroux), some 30 km apart, separated by the pinched synform of Rosis (Ellenberger and Santarelli, 1974). The semi-antiformal structure of Nore, SW of the syncline of Thoré, could also constitute a subdome, roughly 30 km from the main one at Espinouse. Under rigid boundary conditions

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6 **Conclusions**

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this puts d at 15 km = 15×10^5 cm. Another approximation of d is provided by geophysical estimates of the thickness of the crust (30–25 km according to Aubert and Perrier, 1971), diminished by stratigraphical minimum-estimates of the thickness of the supracrustal cover series, at $\sim$6 km, (Arthaud, 1970) or by depth estimates of metamorphism at the core-envelope boundary ($\sim$15 km).

The Rayleigh number, between 65,600 and 248,000, corresponding with a thickness of the source layer of 15–20 km, exceeds the critical value by nearly a full order of magnitude more than does the corresponding number of the Lepontine gneiss dome. This might explain why, in the core of the Agout dome, structural disturbance, homogenization and partial melting have proceeded further and in a more generally distributed fashion than in the Lepontine dome (figs. 1 and 3).

6 Conclusions

Thermally mantled gneiss domes, such as the Agout dome of the French Central Massif and the Lepontine gneiss dome of the Swiss Pennine Alps, possess characteristic features that cannot be fitted to purely conductive models of crustal heat flow. The conductive/convective model of Talbot (1971) fits the observations much better. Of the parameters involved in determining the Rayleigh number, the thickness and the viscosity of the source layer are most important, but – generally speaking – also most difficult to evaluate correctly. Assuming a viscosity in the order of 10^{17} stokes for felsic gneisses at about half their melting temperatures, the thickness of the source layer should be at least 10 km to produce marginal instability and non-cyclic, laminar convection under average geothermal gradients of the continental crust. This could imply that crustal thickening by accumulation of basement nappes of the Penninic type is a pre-requisite for lower crustal convection. Indeed such nappes occur most frequently in the vicinity of thermally mantled gneiss domes, and they invariably pre-date the formation of the domes and their thermal aureoles. Once sufficiently thickened for subsolidus convection to occur, the gneiss layer becomes increasingly liable to thermal instability during subsequent orogenic cycles, while the thickness may also be assumed to be largely responsible for the cyclicity and turbulence of the convecting process. Thus both the extent of granitization and the metamorphic facies series of an orogenic belt (Abukuma versus Barrovian) may be determined largely by the thickness of the orogenic crust. The slightly oblong plan of most thermally mantled gneiss domes suggests that heating from below (by a mantle diapir?) was an equally important source as heating from within the crust (by radio-active decay).

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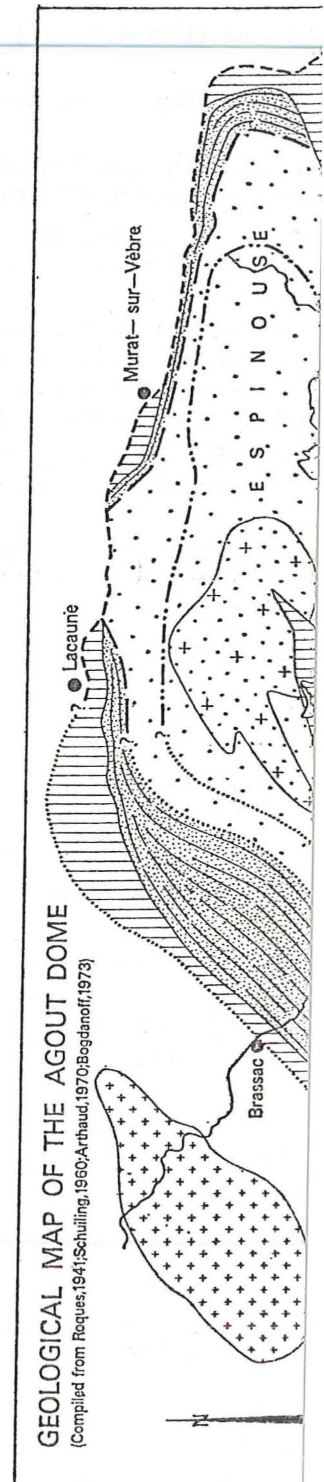

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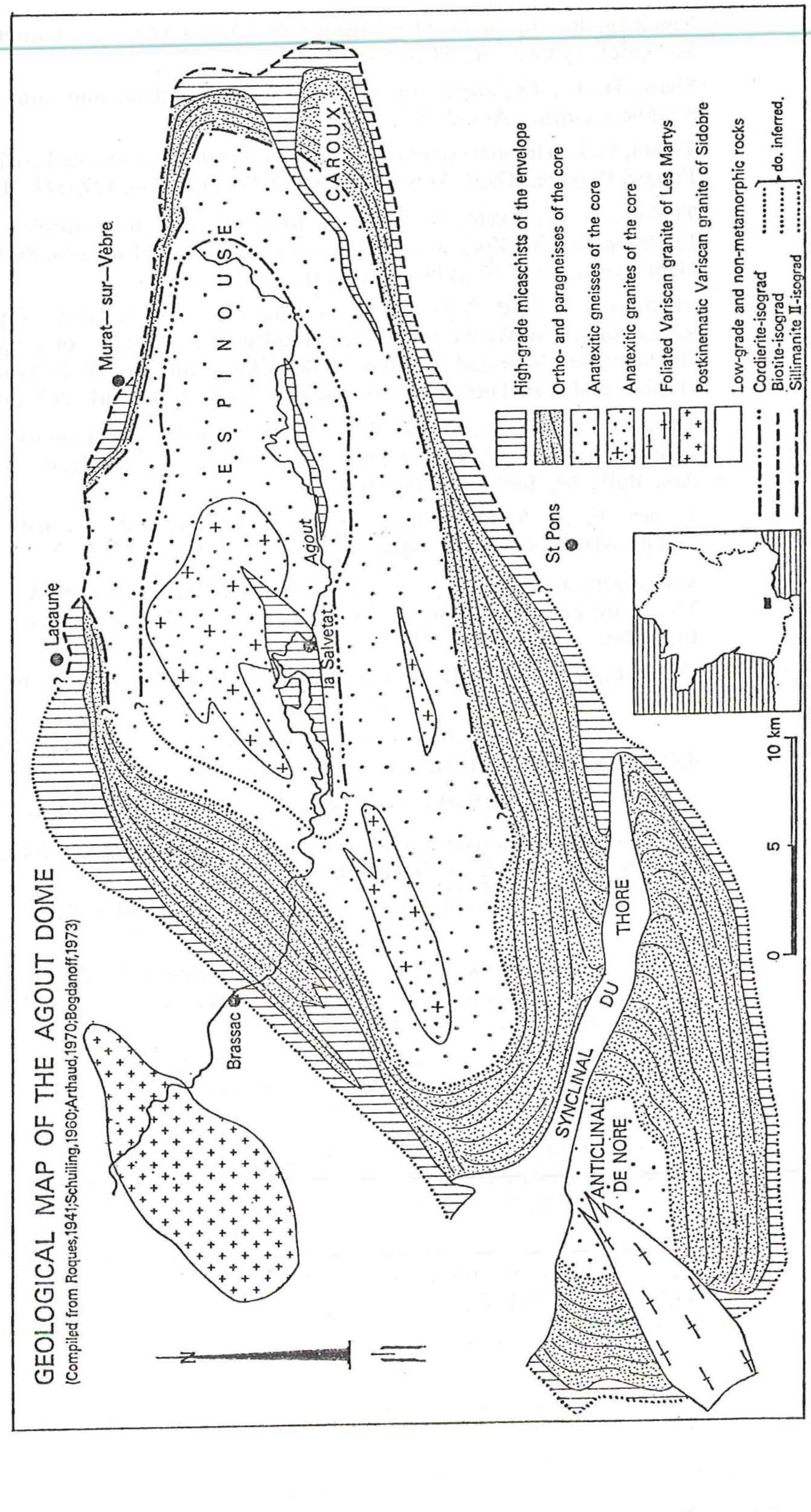


Fig. 1. Geological map of the Agout dome in the 'Montagne Noire' of Central France, showing some of the metamorphic isograds and the probable subdome structures. After Roques (1941), Schuiling (1960), Arthaud (1970) and Bogdanoff (1973).

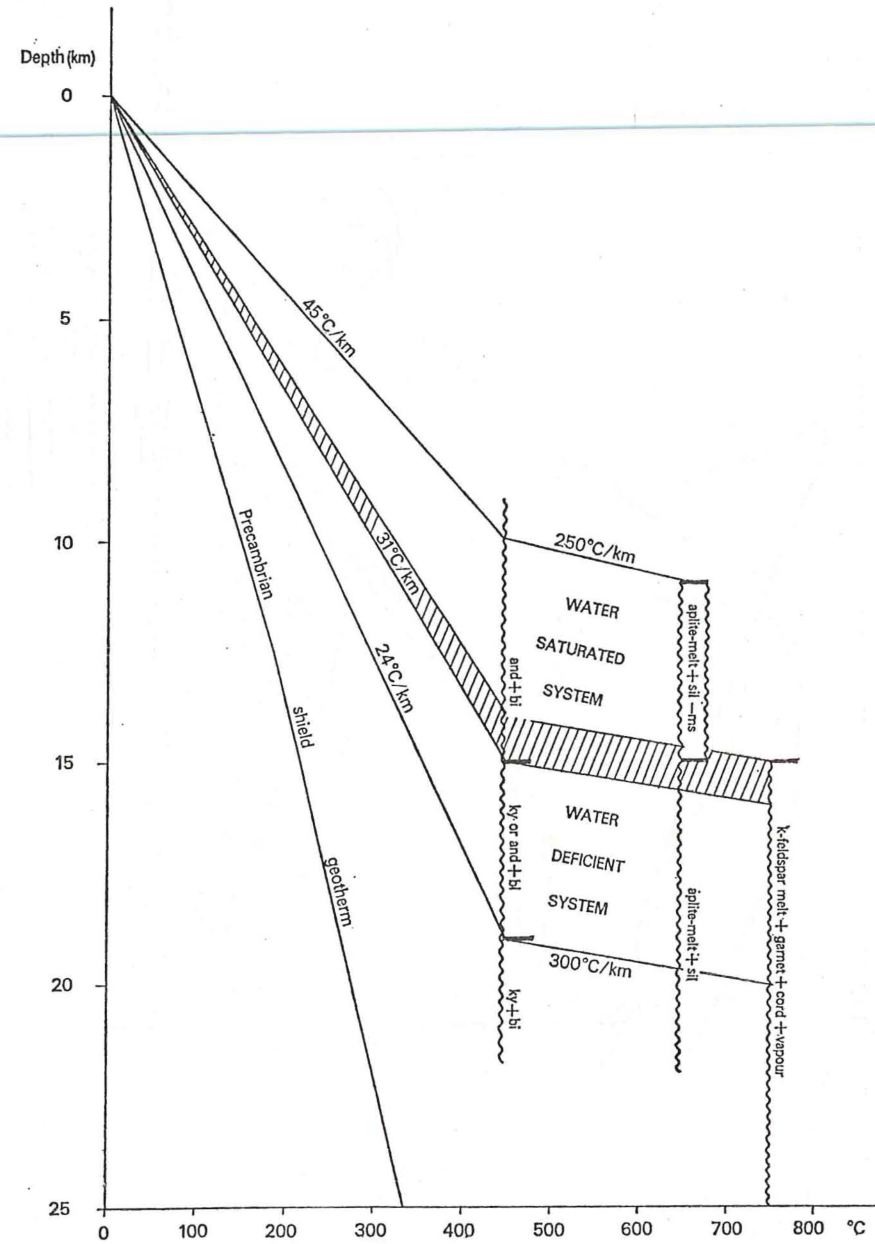
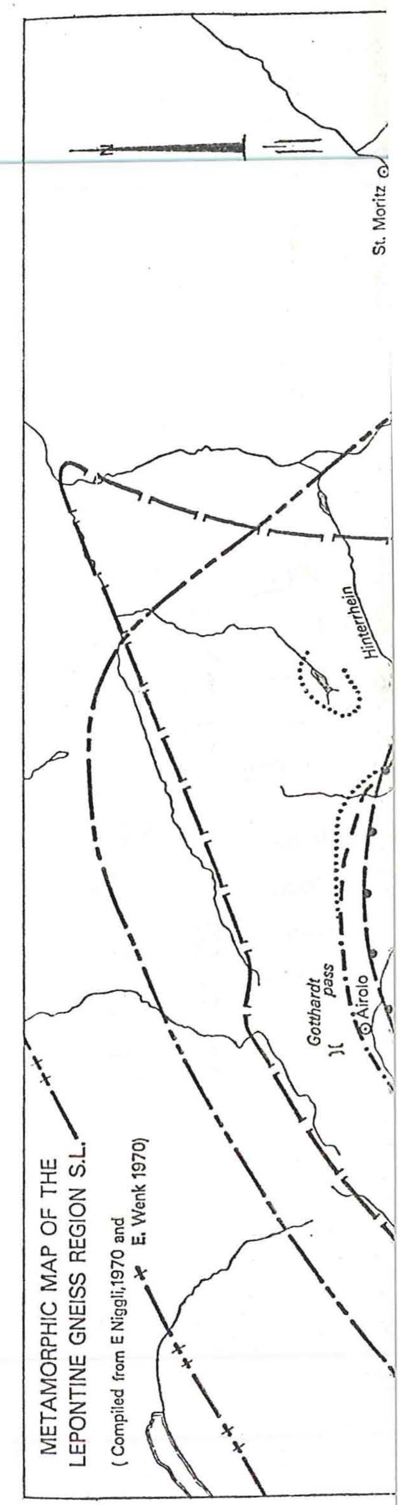


Fig. 2. Range of possible geothermal gradients in the Agout dome and its metamorphic envelope. Average Precambriam shield geotherm shown for comparison. For explanation see text, p. 64–65.

Fig. 3. Metamorphic and plutonic map of the Lepontine gneiss dome in the Ticino Alps, showing some of the metamorphic isograds and the development of alpine pegmaplites, migmatites and granitic rocks. After E. Wenk (1970) and Niggli (1970).



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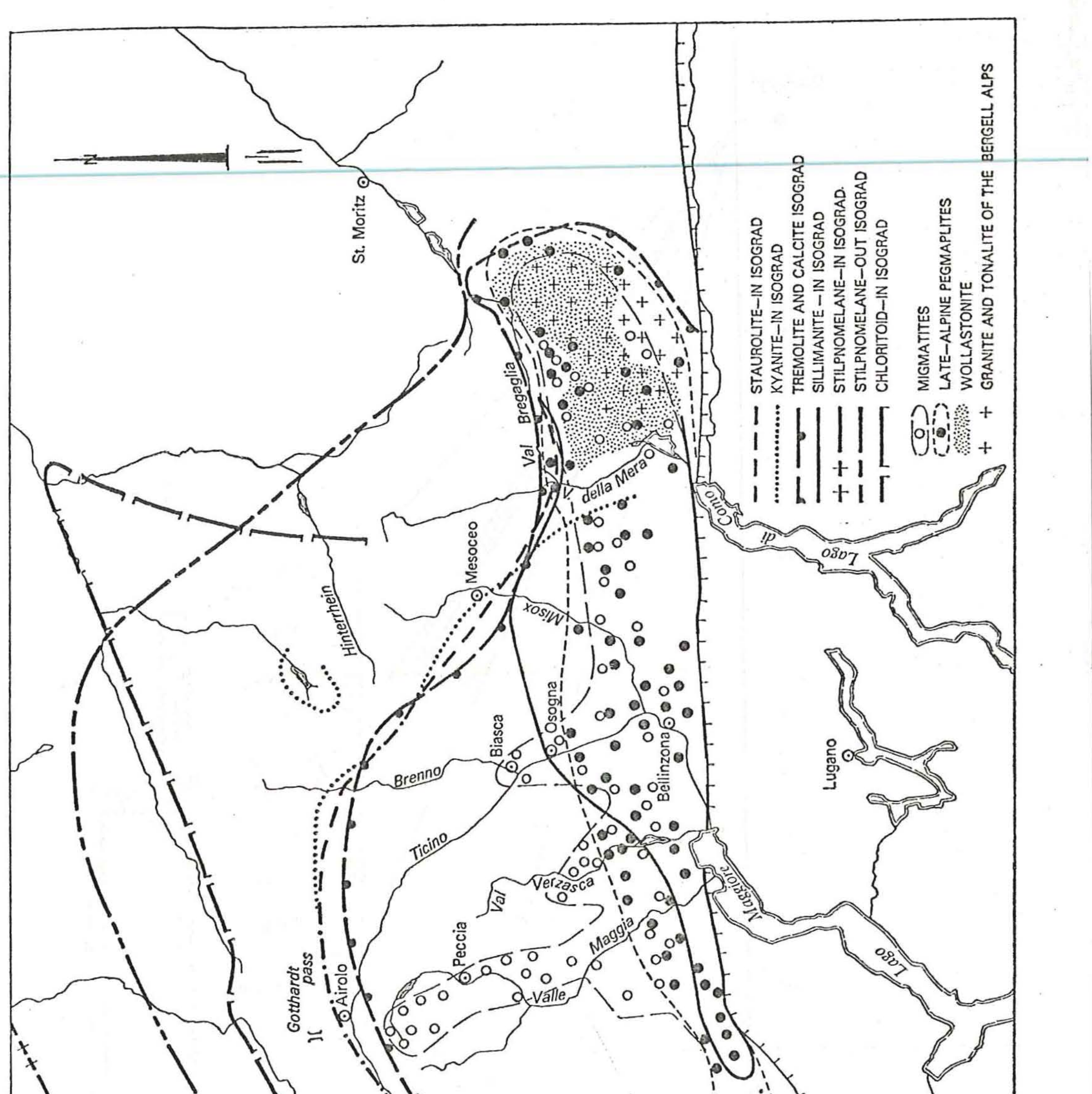

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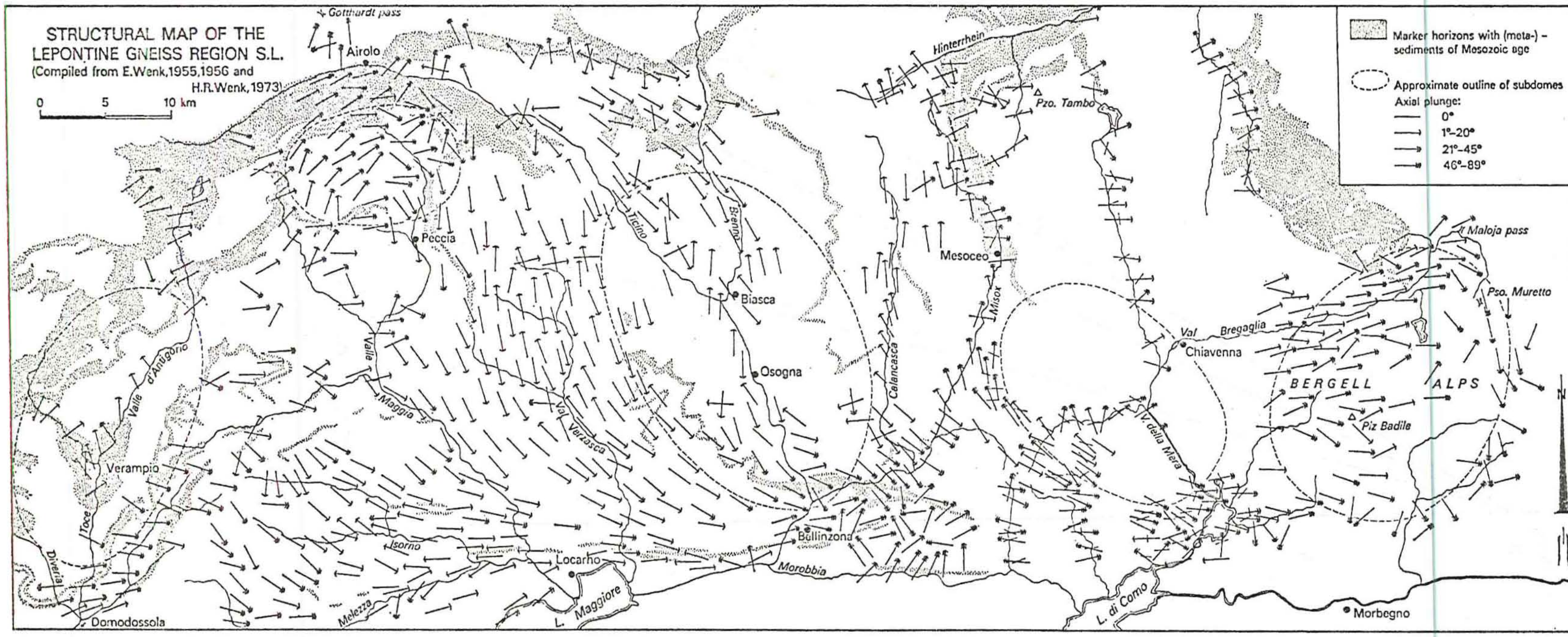
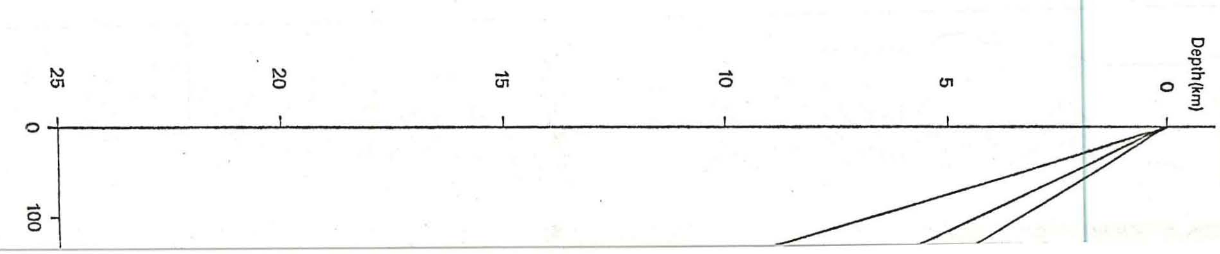


Fig. 4. Lineation map of the Lepontine gneiss region in the Ticino Alps. Approximate locations of subdomes shown with dashed contours. The Maggia cross-zone links the Ticino-culmination (Osogna-Biasca) with a probable subdome in the upper Maggia- and Peccia valleys. Data from E. Wenk (1955, 1956) and H. R. Wenk (1973).



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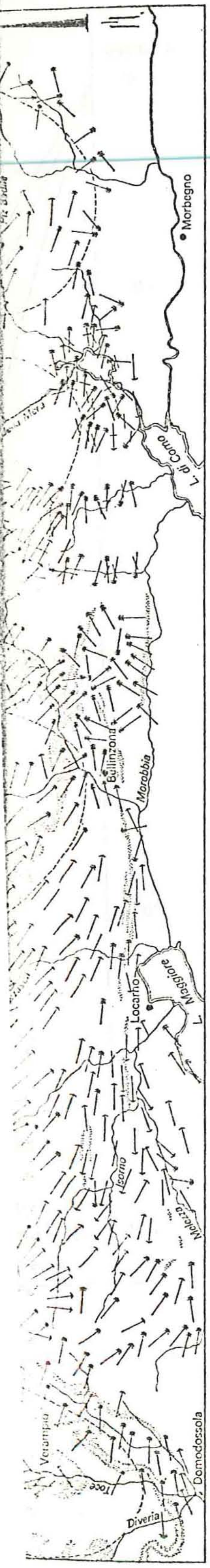

Fig. 4. Lineation map of the Lepontine gneiss region in the Ticino Alps. Approximate locations of subdomes shown with dashed contours. The Maggia cross-zone links the Ticino-culmination (Osogna-Biasca) with a probable subdome in the upper Maggia- and Peccia valleys. Data from E. Wenk (1955, 1956) and H. R. Wenk (1973).

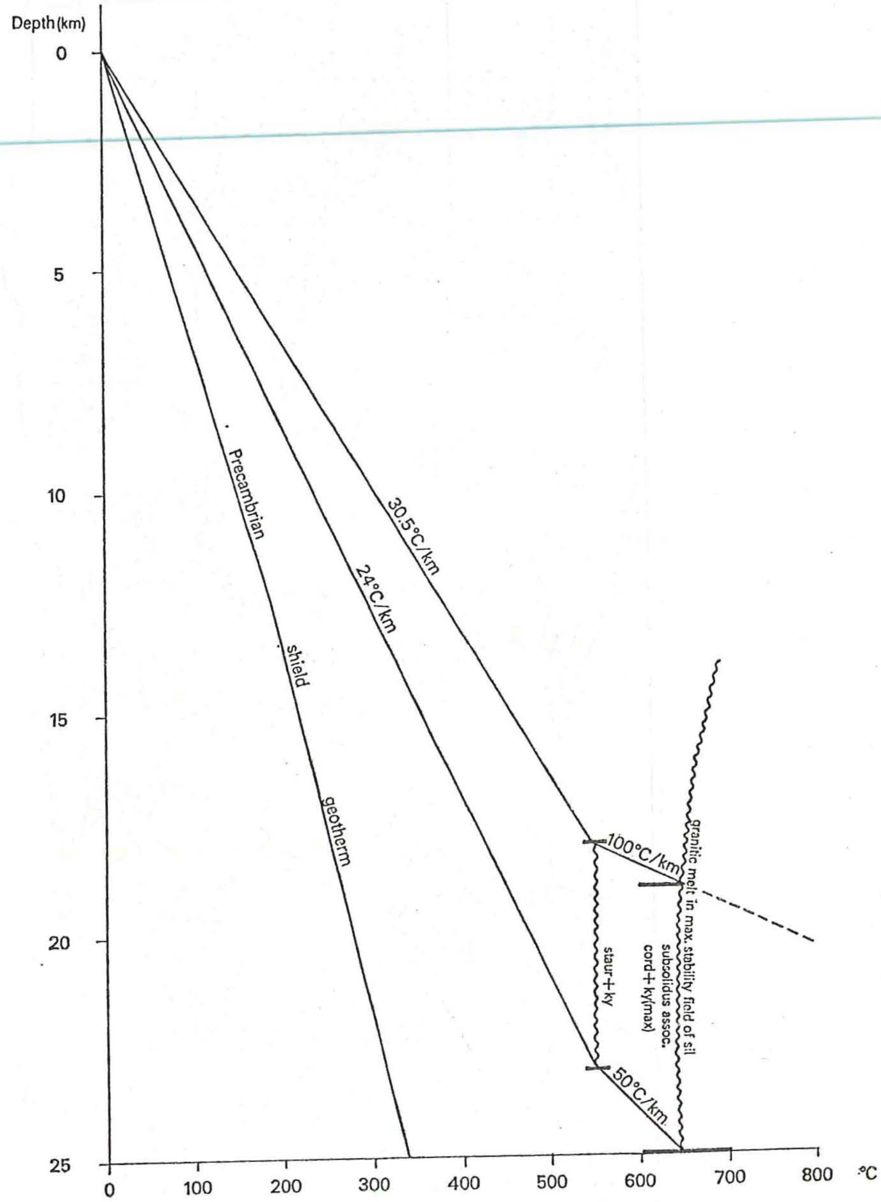

Fig. 5. Range of possible geothermal gradients in the Lepontine gneiss dome and its metamorphic envelope. Average Precambrian shield geotherm shown for comparison. For explanation see text, p. 66.