TECTONIC INTERPRETATION OF REGIONAL CONDUCTIVITY ANOMALIES

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(Accepted 30 April 1993)

Abstract. This review paper selects key results from electromagnetic induction studies of a variety of distinctive tectonic phenomena in the top 200 km of the Earth. Its main theme is that electromagnetic data are essential for an understanding of tectonism involving partial melting, recycling of large volumes of fluids (CO_2 and H_2O) and underthrusting of metasedimentary rocks. The wide variety of tectonic regimes in which these processes are known to be important is reflected in the choice of case studies. A discussion of conductivity models for young oceanic lithosphere and asthenosphere is followed by results from induction studies across the S.E. Australian passive margin, the North American active margin, the Ryukyu Island-Arc and the Oregon Cascades continental arc. The importance of partial melting and free fluid movement i apparent in these regions. Terrain accretion and/or continent-continent collisions recorded at palaeosuture zones in Ireland, Germany and Scandinavia have left distinctive conductivity structures. These are often associated with grain-boundary graphite either in weakly-metamorphosed black shales in underthrust sedimentary basins or precipitated from CO_2-rich fluids. They are discussed in the context of the evolution of mature continental crust. All of the case studies are based on experiments published since 1989 in which the electromagnetic results have been central to an integrated geophysical and geological interpretation.

1. Introduction

My aim in this paper is to present the results of regional electromagnetic (EM) induction studies in a way which stresses their unique ability to contribute to the understanding of some distinctive tectonic phenomena. My approach is one of selecting key results from regions typifying different processes in the plate tectonic cycle and discussing these in the context of pre-existing geological and geophysical data. This occasionally necessitates a brief overview of the tectonic phenomena under discussion. This paper is based mainly on case studies published since 1989, as these tend to be products of the most advanced data acquisition and interpretation techniques. The discussion is restricted to the top 200 km of the Earth and the emphasis is on the investigation of the crust and uppermost mantle. Beyond 200 km depth, current EM induction studies, based primarily on global magnetometer arrays, have not convincingly resolved lateral heterogeneity associated with surface geology.

I begin with the limited success achieved to date in resolving structure beneath young oceanic lithosphere and then move quickly on to more exciting terrain at the ocean-continent transition, in particular the S.E. Australian passive margin and the active margin associated with subduction of the Juan de Fuca plate beneath North America. After a brief summary of modern ideas about the evolution of

Surveys in Geophysics **15**: 123–157, 1994.

continental lithosphere, I go on to describe some of the EM features which I believe are characteristic of this evolution. I start with the formation of nascent continental crust at the Ryukyu Island-Arc and examine the consequences of island-arc accretion to continental crust in the Cascades Range of North America. I consider the structure and composition of the upper crust across a palaeosuture zone in Ireland and discuss the sources of some interesting structures detected by EM methods in the mature crust of Germany. I conclude with EM results from the cratonic Svecofennian Shield which have proved to be so important in our understanding of Precambrian tectonic processes.

The EM induction studies used in this paper are based mainly on magnetotelluric and magnetovariational methods – hereafter referred to as MT and MV respectively. The parameter sensed by these is electrical conductivity measured in Siemens per metre (Sm^{-1}) often displayed in terms of its reciprocal, the electrical resistivity measured in Ohm-metres. Background information on the sources of conductivity variations in the Earth can be obtained in Lastovickova (1991) and Schwarz (1990). The latter review paper and Hjelt (1988) also provide extensive reference lists for regional conductivity studies; these have been updated for the period 1989–1992 in an appendix to this paper.

2. Evolution of Oceanic Lithosphere

Oceanic lithosphere is formed near mid-ocean ridges by the complex interaction of tectonic. petrological and hydrothermal processes. The ridges are extensional zones where two ocean lithospheric plates diverge and upwelling fertile peridotite (lherzolite) mantle undergoes adiabatic decompression and partial melting. The mafic (basaltic) melt and depleted peridotite (harzburgite) mantle remain in equilibrium at depths $> \sim 15$–40 km, above which the rising mafic magma undergoes fractionation. Typical oceanic crust so formed consists of ultramafic/mafic cumulates overlain by sheeted dykes of gabbros followed by tholeitic pillow basalts covered with ocean sediment. Beneath substantial portions of the ridge, it is thought that narrow (<6 km) magma chambers are present in the lower crust, cooled by hydrothermal circulation (eg. Detrick *et al.*, 1987) though some models suggest a broad (>50 km) melt region with a triangular cross-section.

To a large extent, the composition and structure of the oceanic lithosphere is set at a mid-ocean ridge and is only occasionally modified by off-axis magmatic processes. Most of the cooling at the ridge axis is a consequence of hydrothermal circulation but conductive cooling predominates by the time oceanic lithosphere is ~25 My old. The contribution of off-axis hydrothermal circulation to the alteration of oceanic crust is poorly understood. Conductive cooling models however have successfully predicted the changes in ocean-floor bathymetry and heat flow away from the ridge and other temperature-sensitive properties of the oceanic lithosphere are expected to change as it cools. In particular, shear waves studies demonstrate that the depth to the top of a low velocity zone (LVZ) increases with distance from the ridge axis from a depth of ~20 km to a maximum of >100 km. Seismologists believe that the LVZ is caused predominantly by the onset of ~1% partial melting of olivine-rich peridotite, which suggests either the presence of volatiles (e.g. water) to depress the mantle solidus or a high temperature mantle. (Anderson, 1989).

2.1. Young oceanic lithosphere

In comparison with other geophysical results, the electromagnetic view of young oceanic lithosphere is severely limited by a scarcity of data (see review by Constable, 1990). Nevertheless, it is expected that electrical conductivity will be a sensitive indicator to changes in heat and fluid transport processes and preliminary results from Deep Sea Drilling Project data do confirm this. It has already been found that the conductivity decreases with increasing depth and age within the oceanic crust as a consequence of reduced interstitial seawater temperatures and restricted hydrothermal circulation. The latter is probably due to crack sealing by chemical precipitation (Chave *et al.*, 1990). In addition, controlled-source EM experiments (Cox *et al.*, 1986), though limited in number, have shown that the conductivity progressively decreases with depth to values ~10^{-3} Sm^{-1} in the sheeted dyke layer of the crust. They also show that by the time the oceanic lithosphere is 25 My old, the uppermost 30 km of oceanic lithospheric mantle is highly resistive (10^{-5} Sm^{-1}), consistent with a dramatic reduction in water content below the Moho. The oceanic Moho may in fact be a serpentinisation boundary (Anderson, 1989).

The interpretation of oceanic upper mantle conductivities below about 40 km has been and remains controversial. There are problems connected both with the construction of a set of conductivity models consistent with the MT/MV data, and with the petrological consequences of the preferred conductivity distribution. Oldenburg (1981) was the first to present a set of models from seafloor MT/MV data which suggested that a highly conducting zone associated with the seismic LVZ appeared to deepen with increasing lithospheric age. Subsequent interpretations (Oldenburg *et al.*, 1984; Heinson & Constable, 1992) have cast some doubt on the reliability of his original conclusion but both agree that it is still possible for a conducting layer to deepen with increasing age of the lithosphere. One explanation for the source of this zone,consistent with seismic data, is the presence of about 1–3% partial melting of a water-undersaturated (~0.1% H_2O) peridotitic upper mantle (e.g. Tarits & Jouanne, 1990).

However there are alternative interpretations of the EM data which can give rise to a radically different conductivity model for the oceanic lithosphere. Heinson & Constable (1992) have constructed an age-dependent conductivity structure for the oceanic upper mantle above ~200 km based upon existing geothermal models (Parson & Sclater, 1977; Stacey, 1977), a petrological model (Ringwood, 1982) and laboratory determinations of the temperature-controlled conductivity of a dry olivine-rich mantle rock (Constable *et al.*, 1992). Their model (Heinson & Constable, Figure 5) has conductivities two orders of magnitudes lower than previous estimates and they suggest that this is because previous seafloor MT interpretations have not included the distorting effect of coastlines. Their approach has been criticised by Tarits *et al.* (1992) who argue that they have included the effect of mantle volatiles only to the extent of introducing a wet mantle solidus for the standard pyrolite model. Tarits *et al.* also question the validity of adding coastlines to a resistive mantle model in order to satisfy the MT data. In a reply to Tarits *et al.*, Constable & Heinson (1993) vigorously defend their view that a 3d coastline effect is associated with a resistive upper mantle. If Tarits *et al.* are correct in arguing that the upper oceanic mantle is not so resistive – as other studies with a 2d oceanic distortion removed suggest (e.g. Bahr & Filloux, 1989; Wannamaker *et al.*, 1989b) – then the effect of sub-solidus mantle volatiles must be taken into account. Several mechanisms have been postulated e.g. water (Tozer, 1981; Tarits, 1986), grain boundary carbon (Duba & Shankland, 1982) or free hydrogen ions (Karato, 1990), but the current deficiency in laboratory data together with the limited ocean-floor EM data have so far precluded any major EM contribution to our understanding of the evolution of oceanic lithosphere.

2.2. Ocean – continent passive margin

A passive margin is the transition zone between continental and oceanic lithosphere where tectonic activity has long since ceased. Initial extension of continental lithosphere may give rise to crustal thinning and rifting accompanied by the upwelling of hot asthenospheric mantle (e.g. McKenzie, 1978). As extension proceeds, the hot asthenospheric material generates basaltic magma which intrudes or underplates the lower continental crust (e.g. White *et al.*, 1987) making it difficult to locate precisely the oceanic-continental crustal boundary. As the two edges of the old continent move apart they subside because of thermal cooling and subsequent sediment loading from the continent adjacent to each edge. The transition from oceanic to continental lithosphere at greater depths is not well understood. Shear wave tomographic studies have suggested that between about 200–400 km depth, continental shield mantle velocities are slightly higher than oceanic mantle velocities. This could imply a continental tectosphere up to 400 km thick with a 200 km root whose composition and temperature is different from adjacent oceanic upper mantle (Jordan, 1981). Alternatively, this root may simply be convecting upper mantle asthenosphere – possibly colder and weakly coupled to the overlying plate – with no significant compositional differences (Anderson, 1989).

A model for the electrical conductivity structure (Figure 1) across a passive margin has been presented by Kellett et al. (1991), based upon land and sea floor MT data across the southeast Australian continental margin.

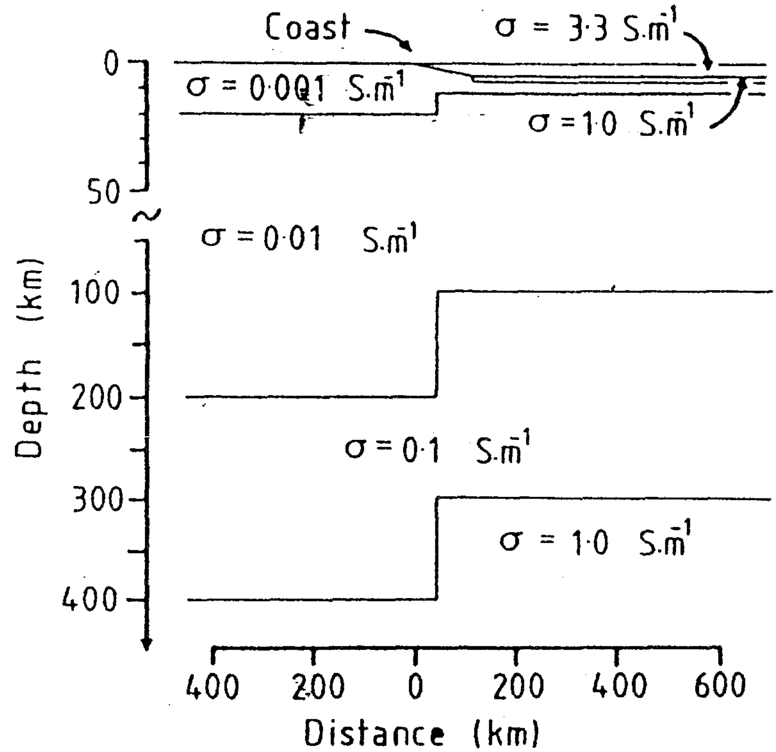

Figure 1. Preferred electrical conductivity model across SE Australia and the Tasman Sea based upon ~20 land and seafloor MT sites operating in the period range 600 to 200,000 s. Seawater: $3.3\,S\,m^{-1}$; Tasman Sea sediments: $1.0\,S\,m^{-1}$; Oceanic and upper continental crust: $0.001\,Sm^{-1}$; Lower continental crust and lithospheric mantle $0.01\,S\,m^{-1}$. Reproduced from Figure 6 in Kellett *et al.*, *Tectonophysics*, 192 (1991), pp. 367–382.

The continental conductivity-depth structure, based upon results from previous experiments, was kept fixed during a systematic search for 2d conductivity models which satisfied the data to within a specified tolerance. Although the parameterisation of the models was coarse, the search confirmed a shoaling of the good conductor ($0.1\,Sm^{-1}$) beneath oceanic lithosphere. This conductor could not be deeper than 200 km beneath the ocean nor could it be immediately below the crust. A shoaling of the highly conducting ($1\,Sm^{-1}$) base of the model on the oceanward side was also suggested but not demanded by the data and is similar in structure to shear wave velocity models for the region. A further search designed to estimate the lateral resolution of the model showed that the vertical boundaries to a depth of 100 km could be moved up to 100 km on either side of the continental shelf without affecting the misfit between data and model response. A similar conclusion can be inferred from a model for the transition zone between continent and quasi-oceanic lithosphere in the Black Sea region (Korotaev *et al.*, 1990).

Kellett *et al.* provide a speculative but plausible tectonic framework for their electrical conductivity model. The continental margin of SE Australia was formed 100 My ago by continental lithospheric extension. As sea floor spreading progressed, the margin moved away from a quasi-ocean ridge spreading centre above shallow hot asthenosphere. This asthenosphere cooled and subsided causing overlying lithosphere to thicken. The present-day asthenosphere is equated with the $0.1\,\mathrm{Sm}^{-1}$ conductor; under the continent it has returned to its pre-rift depth of 200 km and under the ocean basins it has reached 100 km implying a rate of descent of 1 km/My. There is a need to test these conclusions with similar experiments across other passive margins.

2.3. Ocean-continent active margin

At active margins, oceanic lithosphere is thrust (subducted) under a continent or island-arc. Uyeda (1984) has reviewed some of the major processes occurring at active margins. Mechanical coupling between the subducting and overriding plates is controlled by global plate driving forces and the relative motion between the overriding continental plate and the ocean trench above the subducting ocean lithosphere. If coupling is very strong, the ocean lithosphere dips gently down. Substantial portions of pelagic sediments and sometimes oceanic crustal fragments accumulate as thrust-bounded imbricate slices in an accretionary prism on the landward side of a shallow trench (Chilean-type margin). The subduction of some pelagic sediments, oceanic crustal serpentinites, overlying continental crust and mantle provides a wide variety of source materials and volatile components which produce characteristic calc-alkaline andesitic volcanism. If the coupling between the two plates is very weak, accretionary prisms are less developed as more sediments are transported with the downgoing slab in a subduction channel (Shreve and Cloos, 1986). Basaltic volcanism occurs at a continental margin arc landward of a deep trench and ocean ridge basalts may be generated behind the arc where extensional stresses associated with backarc spreading can cause the overriding plate to retreat from the subducting ocean lithosphere (Marianas – type margin). Old, cool oceanic lithosphere, perhaps with an eclogitic upper mantle denser than underlying asthenosphere (Anderson, 1989) is also common at steeply-dipping, weakly-coupled active margins.

The EMSLAB project (Wannamaker *et al.*, 1989a) is probably the most well-known EM experiment across an active margin consisting of MT and MV measurements across the Juan de Fuca subduction zone beneath the west coast of North America. The region has a complex history of terrain accretion (island-arc/continental collision, off-scraping of ocean floor sediments, subduction-related magmatism) and terrain attrition (removal of accreted fragments by subduction erosion, rifting) but more recently its tectonics has been dominated by oblique convergence of the Juan de Fuca oceanic plate beneath the North American continent (Wells, 1990). It is within this geological framework that the electrical resistivity structure across the subduction system was expected to throw light on near surface volcanic

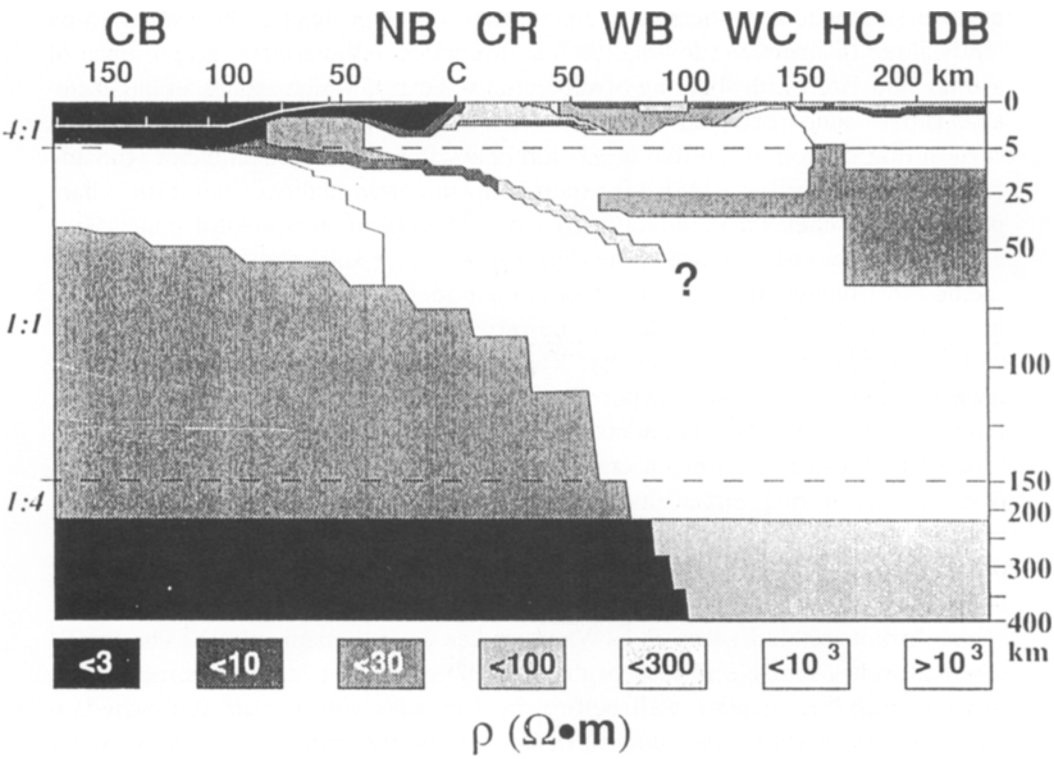

Figure 2. Resistivity cross section interpreted from EMSLAB MT data collected along the Lincoln Line. Section preserves details of the finite element mesh geometry but resistivities have been grouped in half-decade intervals to facilitate presentation. Cascadia Basin (CB); Newport Basin (NB); Coast Range (CR), Willamette Basin (WB), Western Cascades (WC), High Cascades (HC) and Deschutes Basin (DB). From Wannamaker *et al.* (1989b).

and sedimentary structures, the migration of fluid during subduction-related metamorphism and the distribution of magma in the upper mantle. The final model of Wannamaker *et al.* (1989b) for an east-west profile ~400 km long is illustrated in Figure 2.

The most striking feature of the model is a thin conducting (10–100 Ohm-m) layer dipping ~20° beneath the North American continent with conductance (conductivity x layer thickness) decreasing eastward. Following Hyndman (1988), the authors suggest that this conductor represents interconnected saline pore fluids and/or sediments. The lower resistivity of this layer offshore may arise from a small additional contribution from fluids released during mineral dewatering reactions as the subducted slab passes down through blueschist to greenschist and finally amphibolite metamorphic facies. The conductor is similar to the one seen by Kurtz *et al.* (1990) below Vancouver Island, where there is now evidence to suggest that the top of their dipping conductor is coincident with a reflective horizon. This horizon's reflection coefficients of up to ~20% are too high to be explained by fluids alone (Calvert & Clowes, 1990) so the latter authors infer the presence of a zone associated with shearing of subcreted sediments which acts as an impermeable cap to fluids expelled from the oceanic slab below. Shear wave analyses suggest that the top of the slab is ~10 km below the top of the conductor (Cassidy & Ellis, 1991). For the EMSLAB experiment, the poor quality of the seismic data prevents Wannamakei et al. (1989b) from arriving at unequivocal conclusions about the relationship between the dipping conductor and subducting plate.

The electromagnetic view of the ocean-continent transition at the surface can be interpreted easily by appealing to the variable salinity of pore fluids in volcanics and sediments of variable porosity. The resistivity of the accretionary complex between the Cascadia and Newport Basins is, for example, an order of magnitude greater than surrounding sediments (<3 Ohm-m). Intense compressional deformation of this Miocene-Pliocene accretionary complex has resulted in porosity reductions and dewatering, either along imbricate thrust faults or the decollement zone between the accretionary wedge and subducting plate. To the east end of the profile, a low resistivity layer (<30 Ohm-m) is found at lower crustal depths beneath the (forearc) Willamette Basin, more or less coincident with the cessation of subduction related seismicity (Wannamaker & Hohmann, 1991). The implication is that a massive release of fluids from greenschist facies minerals occurs from the subducting plate well before the Cascades volcanic arc is reached, a result consistent with a petrological model offered by Stanley *et al.* (1990a) and discussed later.

The resistivity structure across the ocean-continent transition below ~35 km is more difficult to resolve and is complicated by the presence of a low resistivity (<35 Ohm-m) upper mantle region interpreted to represent upwelling partially melted (<7%) asthenosphere associated with ocean floor spreading at the Juan de Fuca Ridge. Its termination at depth beneath the Western Cascades is consistent with P-wave tomography and the amount of partial melting agrees with that inferred from seismic modelling (but see section 2.1). At depths > ~ 200 km, the model is problematic (Wannamaker & Hohmann, 1991) but even a more reliable interpretation from longer period data (Bahr & Filloux, 1989) can not resolve any variation in resistivity across the ocean-continent transition. This may simply be a consequence of insufficient land data.

3. Evolution of Continental Lithosphere

Continental lithosphere records over periods of several billions of years structural and compositional changes resulting from a variety of tectonic events (e.g. island-arc terrain accretion, continental collision, orogenic collapse, rifting, magmatic underplating). In spite of the high degree of complexity and heterogeneity implied by these processes, geological inferences from the results of deep controlled-source

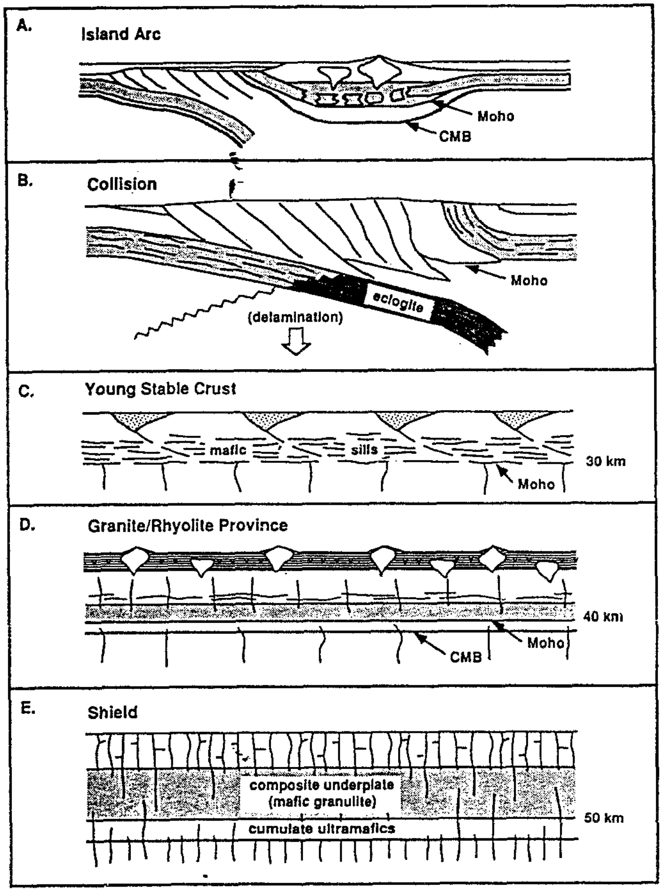

Figure 3. Different stages of continental evolution after Nelson (1991). Stipple – mafic lower crust, irregular black lines in C, D, and E – basaltic dykes and sills, CMB – crust/mantle boundary above which crustal material has developed by differentiation from mantle material below, Moho – seismically defined boundary or steep velocity gradient below which material has P-wave velocity greater than $\sim 8\ \mathrm{km\ s^{-1}}$. Reproduced from Nelson (1991 Figure 1, *Geophysical Journal International*, **105**, pp. 25–35).

seismic studies have provided a framework for understanding its evolution. A schematic diagram illustrating this has been presented by Nelson (1991) and is reproduced in Figure 3.

Figure 3 (A) illustrates an island-arc environment in which pre-existing oceanic crust of basaltic bulk composition is intruded by mantle-derived basaltic magma. The crust/mantle boundary is roughly the boundary between ultramafic cumulate rocks above and residual mantle rock below. The seismically-defined Moho above the base of the crust is probably a transition zone from ultramafic to mafic cumulates (gabbro).

Collision zones develop when island arcs, associated with subducting oceanic lithosphere, or continental lithospheres collide with other continental lithospheres. The arcs accrete to form larger continental land masses often resulting in cratonic cores surrounded by younger crustal terrains. Figure 3(B) shows how a suture zone develops to accommodate the convergence and how large intraplate thrusts allow crustal thickening and mountain building.

Upper lithospheric mantle and parts of the lower crust may either be detached from overlying crust (delamination) or removed advectively by hot asthenospheric mantle. As this sinks, the increase in pressure converts its mafic/ultramafic components into denser eclogite. Its replacement by hot asthenospheric mantle produces a rapid increase in surface elevation and an increase in geothermal gradient often accompanied by crustal melting and magmatism. When the mountain belt is elevated above ~3 km, the gravitational forces associated with overthickened crust are greater than the forces generated by plate convergence and further lithospheric thinning results in orogenic collapse accompanied by mafic magmatism and radial extensional spreading from elevated regions (Dewey, 1988).

The completion of this process can take tens of millions of years and results in crust (Figure 3(C)) whose underlying mantle lithosphere has reached thermal equilibrium. The crust, particularly the brittle upper crust, preserves fault patterns characteristic of the most recent tectonic events. Palaeosuture zones are dominated by seismic reflections dipping in opposite directions (crocodile tectonics; Sadowiak *et al.*, 1991). These reflections can often be correlated with fault zones at the surface and often flatten out at mid-crustal levels. The orogenic collapse associated with palaeosuture zones is sometimes responsible for sedimentary basins and continental rifts, where extensional faults may utilise and/or overprint the fault patterns in the brittle crust developed during suturing.

In addition, deep seismic studies often reveal laminated reflectors in the lower portion of Phanerozoic crust. Nelson (op. cit.) argues that these are due mainly to disseminated mafic sills injected during the replacement of lower crust/upper mantle lithosphere by the underlying hot asthenosphere prior to orogenic collapse. The intermediate (granitic, gneissic or feldspathic) composition of the new lower crust with its basaltic intrusions is petrologically distinct from the olivine-rich mantle below. In addition, the lower crust and uppermost mantle possess quite distinctive temperature-sensitive, depth-dependent rheologies which ensure that the lower crust is more ductile than the mantle. (Meissner et al., 1991).

Cratonisation of young stable crust is envisaged to be the consequence of repeated injection of mafic magmas, evidence of which can be observed in Precambrian shields. (Figure 3(D)) shows how major basaltic underplating may cause anatectic melting of the lower crust resulting in the disruption of laminated lower crust and the initiation of anorogenic silicic (granite/rhyolite) volcanism. The addition of new mafic materials to the lower crust results in a separation and migration downwards of the Moho and crust/mantle boundaries. Whilst acknowledging a multiplicity of views on cratonisation, Nelson (1991) envisages the injection of mafic magmas into stable continents to be an episodic process since the early Proterozoic. In a later section in this paper (Cratonic Continental Crust), I describe work by Korja *et al.* (1992) which suggests that the Fennoscandian Shield was underplated in a time interval restricted to the early-Proterozoic.

The result of magmatic underplating events is tonalite gneiss shield crust (Figure 4(E)) permeated by mafic dykes. The lowermost portion of the crust is a thickened underplate of mafic composition and granulite facies; it gives rise to characteristic P-wave velocities between 7.1 and 7.8 km s^{-1} and a variably reflective Moho which tends not to be such a sharp discontinuity as that observed in younger crust. The crust/mantle boundary between ultramafic cumulates above and mantle below is likely to be some depth beneath the Moho.

3.1. Island-arcs

The northwest Pacific around Japan contains some of the most dramatic examples of island-arc systems associated with subduction zones. Shimakawa and Honkura (1991) have reported the results of a sea floor MV experiment to investigate the conductivity structure beneath the Ryukyu Trench – Arc system. This extends in a NE-SW direction along the margin of the Philippine Sea where oceanic lithosphere is being subducted northwestward beneath continental lithosphere under the East China Sea. A conductivity model for the region (Figure 4) is based upon only five magnetometer stations along a 400 km profile.

The authors assert that only a particular class of models can satisfy their data. This class requires two separate columnar conducting zones (0.5 S m^{-1}); one beneath the Ryukyu Island-Arc and the other beneath the backarc basin of the Okinawa Trough. Each conductor can be moved laterally by 20 km and its conductivity may vary within the limits 0.1 S m^{-1} to 1 S m^{-1}. The island-arc conductor is interpreted to be a plume of water liberated during decomposition of hydrated peridotite, in agreement with a petrological model proposed by Tatsumi (1989). The backarc conductor is thought to be due to partial melting triggered by water released as the oceanic plate is subducted at even greater depths.

While this interpretation is broadly consistent with dehydration and melting models for subduction zones (see Figure 36, Wyllie 1988), it is not unproblematic. Jones (1992) has already pointed out that the high vertical conductance of the forearc conductor implies unrealistically high porosities. Shimakawa and Honkura also admit that the backarc conductor is located too far west of the upper surface of the subducting slab (outlined by earthquake foci) to be consistent with the partial melting model of Tatsumi. Both of these problems may be a consequence of the sparse data set and the inherent lack of vertical resolution in the MV method but, in spite of these, the EM study has given earth scientists their first image of fluid/partial melt processes in an island-arc environment. The backarc conductor here is more likely to be related to backarc rifting beneath the Okinawa

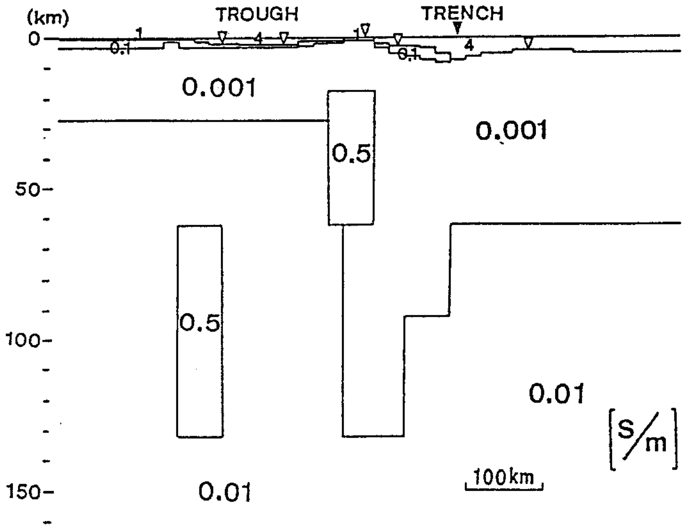

Figure 4. A conductivity cross-section across the Ryukyu Trench – Arc system based upon five magnetometer sites (open triangles). The oceanic lithosphere of the Philippine Sea is assumed to have a conductivity of 0.001 S m^{-1} to a depth of ~65 km. The Ryukyu island-arc lies between the Ryukyu Trench to the southeast and the Okinawa Trough to the northwest. The continental lithosphere beneath the backarc basin of the Okinawa Trough consists of a resistive upper crust (conductivity fixed at 0.001 S m^{-1}). Seawater (4 S m^{-1}) and seafloor sediments (0.1 S m^{-1}) are also fixed in the model. With these constraints, the data are consistent with the presence of the 0.5 S m^{-1} conductors. From Shimakawa and Honkura (1991 Figure 4, Journal of Geomagnetism Geoelectricity, **43**, pp. 1–20).

Trough, reminiscent of the conductor beneath the backarc Deschutes Basin (Figure 2).

A preliminary conductivity model for the ocean-ocean subduction zone beneath the Izu – Bonin island-arc has been presented by Toh (1992). A forearc conductor is evident but the asthenosphere beneath the young overriding oceanic plate is too conductive (0.1 S m^{-1}) and too shallow (<70 km) to resolve any conductive zones on the other side of the island-arc.

3.2. Continental margin arcs

The Cascade Range consists of a Tertiary/Quaternary volcanic belt running parallel to the west coast of North America and extending from northern California to the Canadian border. It is an example of an active continental volcanic arc formed by the subduction of the Juan de Fuca oceanic plate beneath North American continental lithosphere. In spite of its formation at a convergent ocean-continent margin, it is undergoing regional extension possibly related to body forces generated during the orogenic collapse of the Basin and Range province immediately to its east.

Stanley *et al.* (1990a) have combined MT, seismic refraction and heat flow results in an attempt to understand the crustal processes at work in the region.

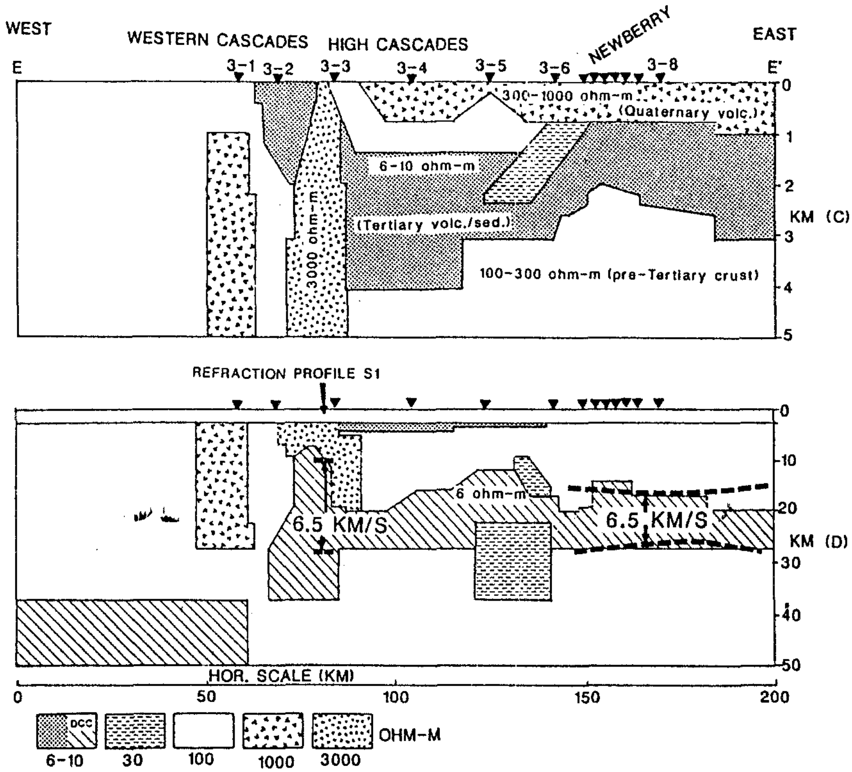

Figure 5. Combined MT and seismic refraction interpretation of Stanley *et al.* (1990a) along their profile EE′ across the Oregon Cascades. The MT stations are marked by inverted triangles. The east end of the line is coincident with a seismic refraction profile and the position of a north-south refraction profile (S1) is marked. The upper crust consists of unaltered Quaternary volcanics where minor pore space containing fresh water gives a relatively resistive layer of 300–1000 Ohm-m. It has a seismic velocity ~2.9 km s^{-1}. The Tertiary volcanics have been weathered and tend to host clays and zeolites which have reduced their resistivity to ~6–10 Ohm-m. This conductive layer is roughly coincident with seismic velocities of ~4.7 km s^{-1}. The pre-Tertiary crust with velocities ~6 – 6.3 km s^{-1} is typical of igneous/metamorphic subvolcanic basement of felsic composition. The excellent correlation in depth between the top of the mid-crustal conductive (6 Ohm-m) layer and the upper boundary of the 6.0 km s^{-1} seismic layer is described in the text. In view of the small number of MT sites, the resistivity model may be slightly overinterpreted.

The Oregon Cascades is the least complex part of the Range and a model for the resistivity structure along a 200km east-west profile is presented above (Figure 5). I will focus on the interpretation of the mid-lower crust.

The conductance of the upper crust is substantially smaller than the conductance of the mid-crustal conductor (~1500 S), suggesting that the latter is well-resolved, in particular the depth to its top surface. High amplitude wide angle reflections along a north-south refraction profile (S1) demonstrate the existence of a sharp transition in velocity to a mid-crustal layer with a well-determined velocity of

6.5 km s^{-1}. This spatial coincidence between the top of this conductor and the upper boundary of the 6.5 km s^{-1} layer is a common feature in the other MT and seismic refraction profiles in and around the Cascade Range. The lower boundary of the conductor can not be resolved by MT as well as the upper one so the coincidence between the base of the conductor and the bottom of the 6.5 km s^{-1} layer may not be reliable. Below the conductor the lower crust has a poorly-resolved resistivity of ~100 Ohm-m and a seismic velocity ~7.4 km s^{-1}. The Moho (velocity >7.9 km s^{-1}) is at a depth of ~35 km.

An interpretation of the deep crustal processes operating in the Cascades is summarised by Stanley *et al.* (1990a) in a diagram reproduced in Figure 6. The mid-crustal conductor is thought to be caused by aqueous fluids generated during prograde dehydration of greenschist – amphibolite rocks or released from crystallising water-rich magmas. A significant contribution to fluids in the crust must also originate from the subducting Juan de Fuca plate giving rise to low resistivity (~100 Ohm-m) lower crust and upper mantle, (see also Figure 2). The wetting characteristics of these CO_2 and H_2O fluids suggest that they are only transported to the lower crust by hydrofracture (Watson and Brenan, 1987). In the crust, high pore pressure fluids may also migrate upwards along grain boundaries (disaggregation). The mafic magmas underplating the crust (7.1 < velocity <7.8 km s^{-1}) generate silicic magmas by secondary melting of overlying rocks. This heat dehydrates greenschist and amphibolite grade rocks producing more hydrous fluids that flux additional silicate melts. Any high pressure pore fluids not combined in these melts may migrate upward via hydrofracture or disaggregation until either enough volume is created to accommodate them or they reach an impermeable zone. This upward migration ceases at the brittle-ductile transition in the Cascades as the depth limit for earthquakes is roughly coincident with the top of the mid-crustal conductor. The reduction in mineral strength due to fluids has been the focus of many laboratory studies in rock deformation (see e.g. Kirby and Scholz, 1984) but it is satisfying that Stanley and his co-workers provide direct field evidence of the association between fluids and ductility in the deep crust.

The combination of MT, seismic and heat flow data has been a powerful tool not only for the interpretation of the Cascade Range but also for an understanding of the active tectonics in western North America. Gough (1992) has focussed upon the role of fluids (water with dissolved CO_2/NaCl and partial melt) in the region and argues for their presence from spatial correlations amongst the different data sets. Thus a shallow high conductivity zone in the upper mantle correlates with a low velocity – high attenuation zone (see also Praus *et al.*, 1990) and high heat flux (see also Adam *et al.*, 1989), whilst high electrical conductivity in the crust is often coincident with high crustal temperatures across different geological provinces. I note that these correlations are often associated with recent or ongoing extensional orogenic collapse, e.g. Omineca Belt of the Canadian Cordillera (Ma-

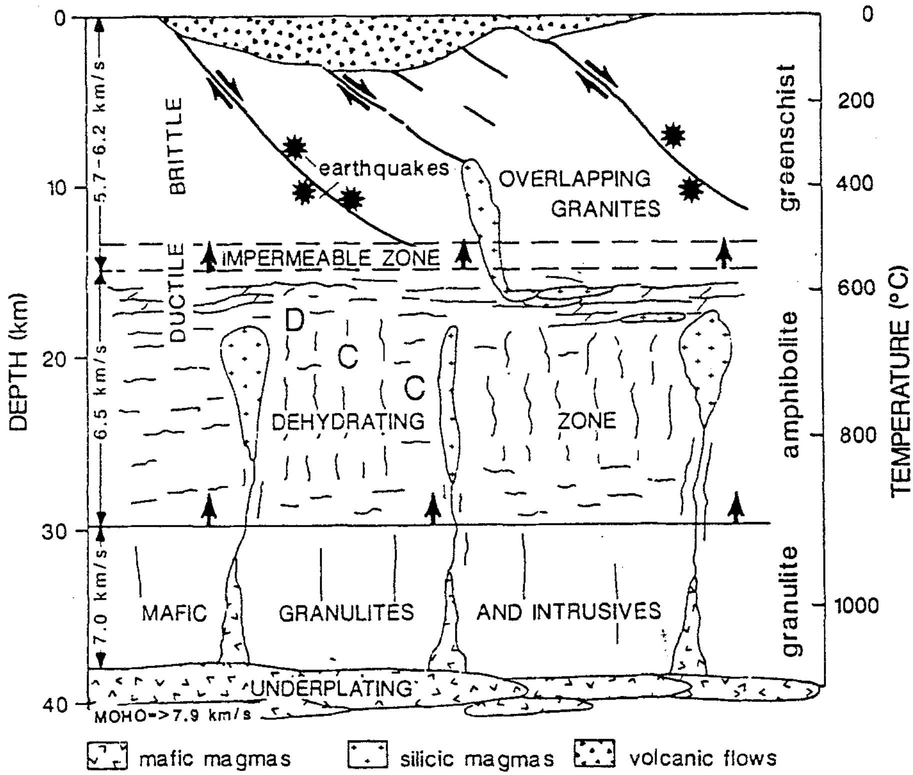

Figure 6. Petrological model of crustal structure of continental magmatic arc, such as the Cascades as inferred from MT, seismic refraction, heat flow and geological data. The general velocity structure from Cascades refraction surveys is showll on the left and the hypothesised geotherm on the temperature scale is shown on the right. Metamorphic facies from the main crustal zones (greenschist – amphibolite – granulite) are also given. The base of the crust is interpreted to be approximately at the dry basalt solidus. Fluids produced as a result of dehydration of the mid-crustal and iower crustal rocks are inferred to rise to an impermeable zone at the top of the mid-crust: the zone of fluids and partial melt is interpreted to produce low resistivities. The heavy arrows imply that as the crust heats up, the greenschist – amphibolite, amphibolite – granulite and brittle – ductile transitions all move upward. After Stanley *et al.* (1990a).

jorowicz and Gough, 1991; Jones et al., 1992a), Basin and Range (Gough *et al.*, 1989), Pannonian Basin (Adam *et al.*, 1989), South Eastern Tibet (Menvielle and Le Mouel, 1990), Central Andes (Schwarz *et al.*, 1992a), and near the Tyrrhenian Basin, in Sardinia (Peruzza *et al.*, 1990).

3.3. Continent – continent palaeosuture zones

The most common features imaged by deep seismic reflection experiments across continent-continent compressional belts are dipping reflectors (Meissner *et al.*, 1991) preserved particularly in the upper and middle crust. Some of these reflectors show correlations with conductors imaged by MT profiles (e.g. Stanley *et al.*, 1990b; Livelybrooks *et al.*, 1992; Korja *et al.*, 1992). Dipping conductors are often a requirement of MT data even when there are no corroborative seismic results (e.g. Adam *et al.*, 1990; Wannamaker and Johnston, 1991; Schwarz *et al.*, 1992b). An example from the Iapetus Suture zone in Ireland (Brown & Whelan, 1994) will illustrate some aspects of the interpretation of conductivity zones preserved in the upper crust.

During the final closure of the Iapetus Ocean, the southern Avalonian continent was subducted north-westward beneath the Laurentian continent giving rise to a complex imbricated accretionary prism partly exposed in Central Ireland as the Longford – Down Massif. This NW – SE convergence was also accompanied by sinistral strike-slip shuffling of a group of terrains along major NE – SW faults. After the convergence ceased, a broad (~150 km) sedimentary basin formed above the suture zone in Central Ireland. This basin suffered differential subsidence along the NE – SW faults which gave rise to a series of NE – SW trending volcanic blocks separated by volcano-sedimentary troughs with widths of ~15 km. Central Ireland is therefore covered with limestones in which occasional inliers of turbidites, greywackes, black shales and island-arc volcanics can be seen.

The resistivity model for a 180 km NW–SE trending MT profile across Central Ireland is shown in Figure 7. In the NW of the profile, a zone of low resistivity (<100 Ohm-m) can be observed between the Ox Mountains (OM) and north of the Curlew Mountains (CM) at a depth of ~10 km. It appears to provide the source of the Fair Head – Clew Bay Linear (FCL), a magnetic lineament of ~150 nT which defines the northern boundary of the North West Terrain in Ireland. The southern boundary of this terrain (Ryan *et al.*, 1993) crosses the resistivity profile near station 4. The model shows a well-resolved lateral change in resistivity across this boundary from <300 Ohm-m beneath the North West Terrain to >3000 Ohm-m beneath the Midland Valley Terrain. The southern boundary of the Midland Valley Terrain is associated with the Southern Uplands Linear (SUL), a well-defined, low-amplitude magnetic lineament trending NE-SW across Ireland. This boundary is also well-resolved by the MT profile particularly in the top 3 km. In the upper crust beneath the Longford – Down Massif, a zone of low resistivity (<10 Ohm-m) can be seen to dip at ~8° southeastward from a depth of ~5 km to a depth of ~9 km. This zone's conductance is ~7,000 S and it is associated with a magnetic anomaly of ~150 nT known as the Virginia Linear (VL).

If this conductor were due only to saline fluids, then such a high conductance

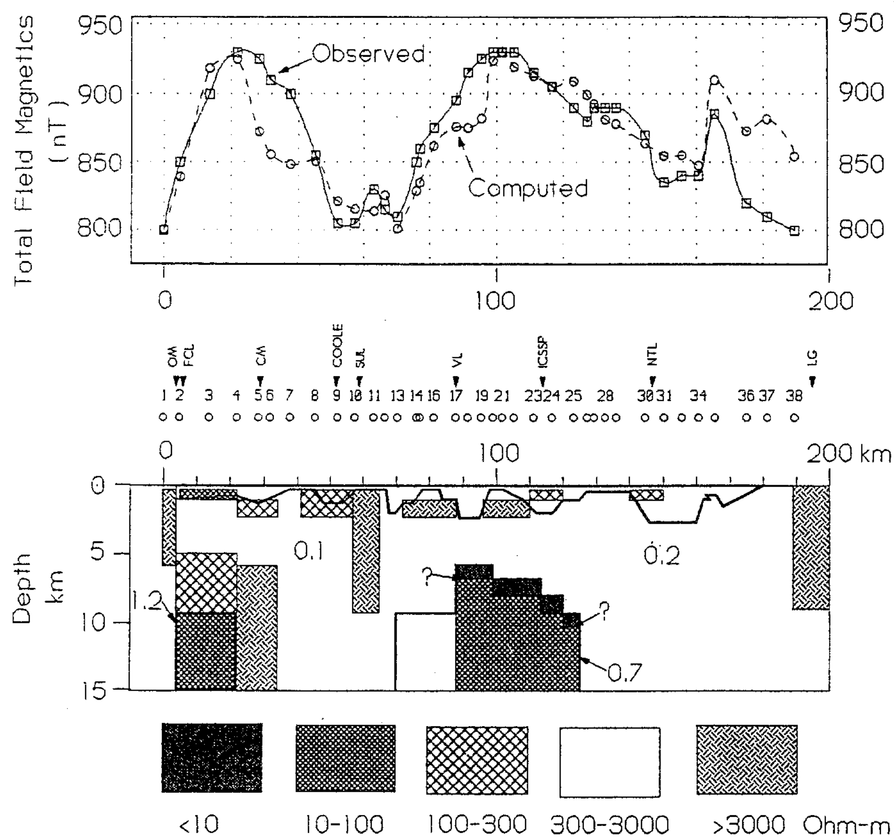

Figure 7. A model for the upper crust across the Iapetus Suture zone in Central Ireland from Brown & Whelan (1994). The block-trough structure at the surface is based on a gravity and magnetic interpretation, subsequently confirmed by MT data. The upper crustal conductor between MT stations 17 to 26 is beneath the Longford-Down Massif. It has a top surface dipping southeastward. The geometry of this surface is reasonably well-constrained but the data cannot resolve structure below it. Numbers above the Earth's surface refer to locations of MT stations. Ox Mountains (OM); Fair Head – Clew Bay Linear (FCL); Curlew Mountains (CM); Southern Uplands Linear (SUL); Virginia Linear (VL); Navan – Tipperary Linear (NTL); Leinster Granite (LG); Solid line: observed magnetic anomaly after removal of the IGRF 1975 and addition of 1000nT; dashed line: computed magnetic anomaly. The magnetic blocks at the base of the model are outlined by bold lines and are roughly coincident with sources of resistivity lower than ~100 Ohm-m. Induced magnetisations given in Amp m^{-1}.

at temperatures ~200 °C requires implausibly large porosities (Jodicke, 1992). There are however two other possible sources – metamorphosed graphitic sediments and/or serpentinised island-arc crust – which are both allowed by the geophysical and geological data.

The preferred source for the conductor is a metasedimentary basin thrust under the leading edge of the southern continent. Stanley (1989) has argued that suture zones in the western Cordillera of the USA and the Carpathian Mountains are characterised by dipping conductive slabs of ~1–5 Ohm-m that can extend to the lower crust. Their low resistivity is due to black shales which are a major component of flysch sequences in basins between two converging continents. The black shale component can have resistivities ~10 Ohm-m or less if they are subject to low-grade (200–400 °C) metamorphism (coalification) to produce metanthracite, whereupon a submicroscopic grain-boundary carbon film establishes a conducting network (Duba *et al.*, 1988; Jodicke, 1992). The black shales may also contain metallic minerals particularly iron which could exist as pyrrhotite (FeS_2) in the sulphur-rich reducing environment of a flysch basin (e.g. Frost, 1991). Typical susceptibilities for pyrrhotite are 0.001–0.1 (SI) which are large enough to provide the magnetisations ~1 Amp m^{-1} required to explain the 150 nT magnetic anomaly.

An alternative source for the low-resistivity/high-magnetisation zone is based upon a tectonic model for the Iapetus Suture zone in Central Ireland (Morris, 1987). This requires a remnant island-arc buried beneath the Longford – Down Massif. Low temperature (<500 °C) hydrothermal alteration (serpentinisation) of the arc produces serpentine-group minerals such as magnetite from the olivines and orthopyroxenes. The process can enhance conductivity via mineral conduction associated with interconnected magnetite and perhaps increased porosity (Stesky & Brace, 1973) giving resistivities < ~ 100 Ohm-m. The (induced) magnetisation of ~1 Amp m^{-1} implies about 1% magnetite (Toft *et al.*, 1990). The model of Morris (ibid) requires the deposition of pelagic graphitic shales which could still provide a substantial contribution to the total conductance. I'he serpentinised arc would provide conductivity and magnetisation sources. A serpentinised arc may also be the source of the coincident conductivity and magnetic anomalies observed in the North West Terrain. In this area, Ryan & Dewey (1991) have developed a tectonic model which also involves burial of an island-arc though Ryan (pers.comm.) does not rule out the possibility of a buried metasedimentary basin.

3.4. Mature continental crust

The European GeoTraverse (Blundell *et al.*, 1992) provides an excellent framework for a discussion on characteristic conductivity anomalies in mature continental crust. Stretching 4600 km from Tunisia to Scandinavia and crossing a wide variety of tectonic environments, the EGT summarises seismic structural information and physical properties of the crust and upper mantle to a depth of ~500 km. I will focus upon crustal conductivity anomalies as one moves from the relatively young (Phanerozoic) crust beneath Germany to the cratonic Fennoscandian Shield.

The central segment of the EGT runs more or less north-south through Germany from the Alps to the Baltic Sea. It crosses four major tectonic units associated with the Variscan orogen in Europe: the Moldanubian, Saxothuringian and Rhenoherynian zones and the North German Lowlands (Figure 8). Berktold (1991) has compiled MT and MV data across the region and has mapped 3 regional conductors, extending roughly east-west for several hundred kilometres. Two conductors correlate extremely well with the suture zones of the Moldanubian – Saxothuringian (MS) and Saxothuringian – Rhenohercynian (SR) terrains respectively. The North German Conductivity Anomaly has its axis of maximum north-south gradient in conductance at position 'L' in Figure 8.

The tectonic significance of these conductors can be best appreciated by referring to the crustal section published by the ERCEUGT Group (1992). The MS and SR conductors at positions F and H respectively have high conductances and appear to be correlated with magnetic anomalies. A graphite source seems most probable. The SR conductor appears to sole southwards into an anisotropic high conductivity layer on top of a low-velocity zone although the data do not require the presence of two separate layers. The south-dipping isotropic conductor ($\sim$0.05 S m^{-1}) beneath the Rhenohercynian zone is a seismically transparent 2 km thick layer bounded by strong seismic reflectors. Volbers *et al.* (1990) have speculated that this could either be interconnected graphite accumulated by solid phase lateral accretion from metasediments into a low pressure reflective shear zone or graphite precipitation along a shear zone from CO_2 – rich fluids from magmatic or biogenic sources. They seem to prefer however an explanation in terms of underthrust metamorphosed black shales (see Continent - Continent Palaeosuture Zones) which could be connected to the early-Palaeozoic black shales responsible for the high conductivity zone beneath the North German Lowlands. Volbers *et al.* (1990) make the point that the low mechanical shear strength of the shales may have provided a gliding (decollement) zone for Variscan thrusting. Adam *et al.* (1992) have made similar suggestions for the Periadriatic lineament where low-strength graphitic zones have facilitated the escape tectonics of the Pannonian Basin.

The conductivity of the lower crust beneath the central segment is difficult to estimate, particularly in the north, because of the screening effect of conducting layers in the upper mid-crust. The ERCEUGT Group imply that lower crustal conductivities in the depth range 15–30 km are greater than 0.01 S m^{-1} but only publicise two discontinuous zones at depths of 20 km beneath the Rhenohercynian whose conductance is high enough to resolve. These are modelled as relatively thin ($\sim$3 km) highly conducting (0.3 S m^{-1}) layers and the authors remark that the southern zone is coincident with a crustal segment which displays increased seismic reflectivity. They refrain from speculation about possible sources for these zones but I offer the following digression merely to give a flavour of the debate on lower crustal conductivity.

There are few disagreements about fluid sources in the active tectonic regions described earlier; the problems arise when seeking explanations for lower crustal conductivity in stable continental crust. A petrological problem is that lower crustal granulite facies rocks should have absorbed free water leaving a dry, insulating lower crust (Yardley, 1986). A problem also occurs when considering mechanisms

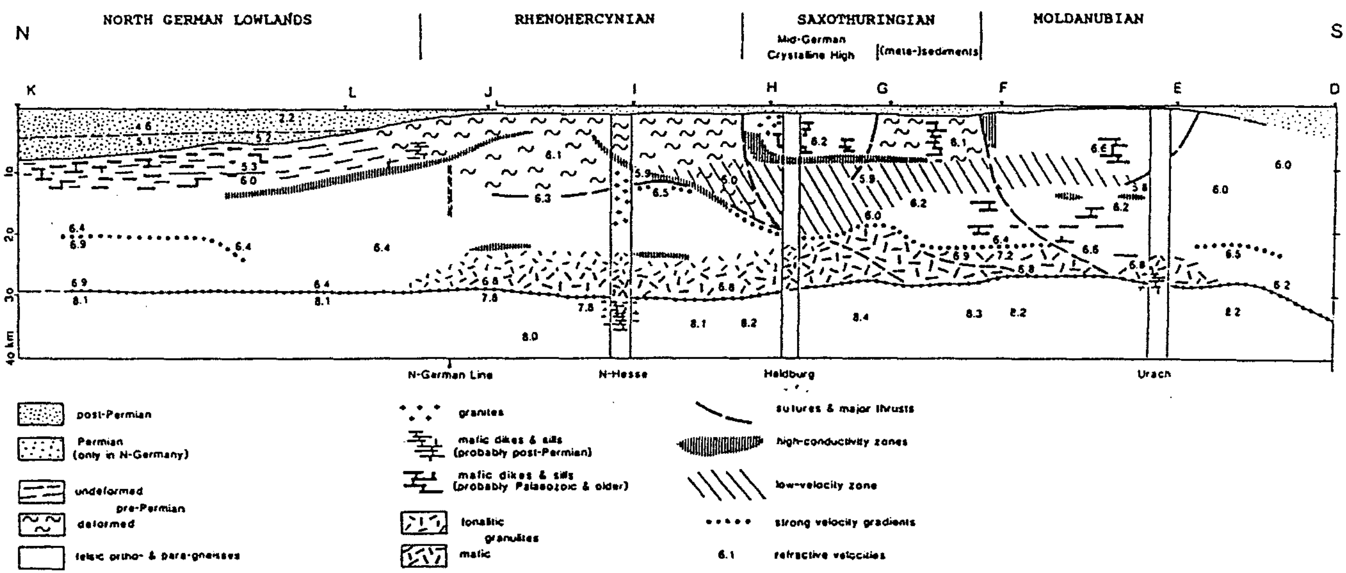

Figure 8. An 800 km crustal section along the central segment of the EGT based on seismic and geological data interpreted by Franke *et al.* (1990). Electrical resistivity results were obtained from an MT profile consisting of several sections at different azimuths which have been projected onto the seismic section by the ERCEUGT Group (1992).

to trap free fluids in the lower crust. Hyndman and Shearer (1989) have argued that interconnecting pore fluid networks are necessarily permeable and the mechanics of fluid transport suggest that free fluids should leak to the Earth's surface in a geologically short time. One way round both problems has been suggested by Bailey (1990) who advocates the accumulation over time of fluids in extensive, thin horizontal reservoirs associated with the brittle-ductile transition zone. This mechanism is difficult to reconcile with the suspected correlation between decreasing seismic velocity and increasing conductivity in the lower crust (Marquis & Hyndman, 1992; Hyndman & Klemperer, 1989) as it is not obvious how the reductions in seismic velocity reported in these studies can be achieved by very thin reservoirs.

A more plausible model for lower crustal conductivity consists of unconnected highly conducting lamellae within a resistive matrix (Merzer & Klemperer, 1992). The motivation for this model comes from the coincidence in some regions between high conductivity and high seismic reflectivity – the latter originating as an interference pattern due to closely-packed lenses of different velocity and density. The conducting lamellae may be due to mantle-sourced volatiles mostly H_2O in mafic sills (see Nelson, 1991 and Figure 3) exsolved as saline fluids; they may be tectonically emplaced paragneisses which have carried connate water down from the upper crust or they may be brine-saturated shear zones where rocks have retrograded to mineral assemblages of greenschist – amphibolite facies in which saline fluids are stable (Sanders, 1991). The resistive matrix may consist of dry kilometre scale lenses of granulite facies rocks whose margins are progressively sealed with the reaction products (e.g. halite) of retrogressive hydration (Sanders, ibid). The particular model presented by Merzer and Klemperer (1992) has several important properties: (1) it predicts the correct orders of magnitude for lower crustal conductivity, (2) it sidesteps arguments against geologically long fluid residence times, (3) it avoids petrological objections to free fluids co-existing with granulite-facies rocks, (4) it could explain lower crustal conductivity anisotropy observed, for example, by Kellett *et al.* (1992) and Tezkan *et al.* (1992) and (5) it could explain the tendency for Phanerozoic lower crust to be conductive and reflective and for cratonic lower crust to be less-conductive, diffractive and have a higher velocity.

It is worth noting that even in active tectonic regions where the evidence for free fluids is most persuasive, high conductivity – lower velocity seismically reflective correlations are not necessarily observed. This may be explained by the effects of velocity anisotropy (e.g. Stanley *et al.*, 1990a) or by the recognition that free fluid is only one of several sources for lower crustal reflectivity (e.g. Jones *et al.*, 1992a and references therein). In mature continental crust however, there is also the possibility that grain-boundary carbon films are responsible for lower crustal conductivity, (Frost *et al.*, 1989). Auger scans on gneisses from the Kapuskasing uplift in Canada (Mareschal *et al.*, 1992) indicate that graphite is a more important contributor to the lower crustal conductivity in the Archaean shield than brines or other solid conductors. Their results suggest that Kapuskasing rocks cooled under Archaean metamorphic conditions that were favourable to graphite precipitation, whereas evidence for saline fluids is sparse. Haak *et al.*, (1991) emphasise that favourable conditions for the subsequent stability of graphite films are controlled by temperature-dependent oxygen fugacity. This also implies the existence of stratified lower crustal conductors. Merzer and Klemper (1992) speculate that CO_2-rich fluids could also preferentially precipitate graphite in sub-horizontal permeable zones so their lower crustal conductivity model does not necessarily allow discrimination between graphite or free fluid sources.

3.5. Cratonic continental crust

The EM data base for the Fennoscandian Shield is one of the most extensive in the world and it is perhaps not surprising that the crust here displays almost the whole range of Earth resistivity (10^{-2} to 10^5 Ohm-m). Some of the key results from MT and MV experiments are given in Korja (1992) and summarised in Figure 9.

The discontinuous conductor from Lake Ladoga via the Kainuu Schist Belt to Oulu, the southern Finland anomaly (Lake Ladoga – Tampere Schist Belt-Pori) and the Kokkola anomaly (Pori – Oulu) surround the homogeneous electrically resistive Central Finland Granitoid Complex (CFGC). The first conductor is thought to represent the boundary between the Archaean and Early Proterozoic Svecofennian Domains while the others may represent boundaries of accreted Proterozoic terrains. These extensively studied conductors are mainly due to graphite- and sulphide-bearing metasediments buried during collisional processes, but saline fluids in fractures also contribute to their low resistivity (<10 Ohm-m). The upper crust (10–14 km) of the CFGC has resistivities in the range 10,000–35,000 Ohm-m approaching that of dry rocks. In this case this is because near-surface groundwater is relatively resistive (>100 Ohm-m) and cannot reduce the bulk resistivity further. The lower crustal conductor (~20 Ohm-m) interpreted to dip northwards beneath the SVEKA profile exists only under the Proterozoic CFGC. There are arguments which suggest – but do not require – that its source is graphite, perhaps deposited during the Svecofennian orogeny (1900–1760 My). The top of the conductor does not follow velocity boundaries observed in seismic refraction data implying that it is an internal terrain boundary (detachment zone) cutting across different crustal lithologies.

The tectonic development of the Fennoscandian Shield has been elaborated by Korja *et al.* (1992) using all the available EM and seismic evidence. In contrast to the generalisation of Nelson (1991), cratonisation of this shield did not take place by episodic mafic underplating to produce a thickened crust with a lowermost high velocity layer (Figure 3(D), (E)). The evidence here suggests that there is great variability in crustal thickness, particularly in the lowermost crustal layer. The thick crust is associated with decreasing seismic reflectivity of the lower crust and

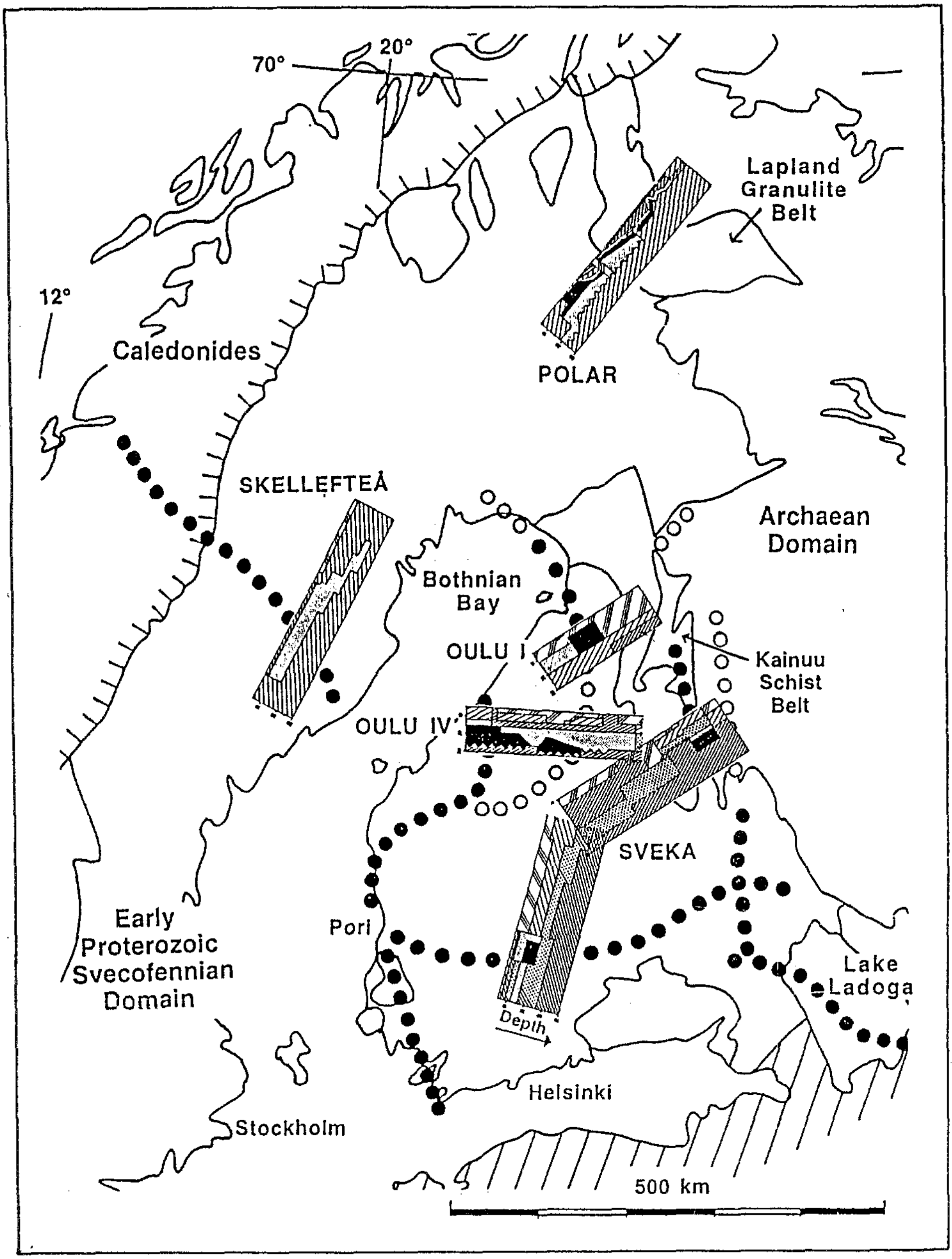

Figure 9. The important conductivity anomalies in the Fennoscandian Shield (from Korja, 1992). Black and open dots denote near-surface positions of anomalies detected by magnetometer arrays. These conductors are usually inclined at depth. The upper surface (depth = 0) of the two dimensional resistivity models corresponds to the MT profile on which the results are based. The vertical axis shows depth in kilometres with tick marks at 10, 30, 50 km etc. The least resistive zones (~1–10 Ohm-m) are coloured black; dotted zones ~10–100 Ohm-m; coarse diagonal lines ~100–1000 Ohm-m; fine diagonal lines ~1000–5000 Ohm-m; banded diagonal lines >5000 Ohm-m.

Moho whilst zones of thin crust have increased reflectivity and a sharply-defined Moho. The authors go on to describe how the roots of collisional belts – defined by their relation to the major palaeosuture conductors – were subsequently thinned during an extensional period. It was this extensional period which gave rise to mafic underplating which, in turn, produced the anorogenic granitoids seen in Fennoscandia. The thickening and thinning of the Fennoscandian crust took place in a restricted time interval (2100 My – 1500 My) and since then the central part of the shield has remained intact. It is worth noting that these important conclusions about Precambrian plate tectonic processes have been made possible only by a careful integration of the EM, seismic and geological results. The EM results were particularly useful in locating the metasedimentary rocks squeezed between colliding crustal blocks and in some instances in defining internal terrain boundaries that extend at least to the middle crust.

4. Concluding Remarks

I have demonstrated that a wide variety of active tectonic phenomena near the Earth's surface have distinctive conductivity signatures usually associated with partial melting and/or the presence of free fluids (water with dissolved NaCl and CO_2). Based on the case studies presented in this paper, I conclude my review with a suggested agenda for future EM induction experiments which. I believe will contribute substantially to our understanding of tectonic processes.

Beneath *young oceanic lithosphere*, more data is required to examine hydrothermal circulation in the oceanic crust and the presence or otherwise of small scale mantle heterogeneities (e.g. magma chambers) thought to exist beneath mid-ocean ridges (Marsh, 1987). The acquisition and interpretation of such data is difficult (e.g. Evans *et al.*, 1991) but several research programs are already being pursued (e.g. Heinson *et al.*, 1992; Chave *et al.*, 1992; Evans *et al.*, 1993).

The precise relationship between oceanic upper mantle conductivity and partial melting requires clarification and one approach may be to extend the geographic coverage of EM data across different *ocean-continent passive margins*. Young passive margins are likely to yield useful results as the oceanic asthenosphere is probably shallow and will give rise to other pronounced geophysical and geochemical anomalies. An additional EM contribution may be to establish whether there is a conductivity signature associated with the continental slope that can be correlated between the two passive margins on either side of an ocean. Characteristic seismic reflectors and magnetic anomalies have been reported (e.g. Talwani, 1989) at a number of passive margins; is it possible to obtain an EM equivalent?

The EMSLAB and Cascade Range studies have demonstrated how EM induction experiments have been central in quantifying fluid/partial-melt processes at *ocean-continent active margins* and *continental margin arcs*. The contribution of these processes to the evolution of the continental crust should not be underestimated; they have an enormous influence on crustal rheology (Meissner & Wever, 1992), mass transport (Torgersen, 1990) and regional metamorphism (Wyllie, 1988). The last author also comments that 'unravelling the processes during subduction remains one of the major problems for understanding mantle heterogeneities and the evolution of the continents'. A priority for ocean-based EM studies must therefore be further experiments at different active margins and in particular at *ocean-ocean subduction zones* to confirm or otherwise the columnar conductors observed in island arc environments. This will require dense networks of seabed MT instruments and considerable input from other earth science disciplines.

There is also a noticeable lack of reliable EM induction data in converging continent-continent environments. Schreyer *et al.* (1987) have suggested that magma generation in these collision zones is strongly influenced by fluids within continental crust that can be subducted to great depths within the mantle. Given the recently published results for the European GeoTraverse (Blundell *et al.*, 1992), the European Alps may offer some advantages over other collision zones. Controlled-source EM instrumentation to overcome noise at short periods, tensor decomposition methods to minimise distortion and probably 3d interpretation will be required. Preliminary results for regional differences in the thickness of the electrically-resistive lithosphere have already been presented (Bahr, 1992), but there is little information on crustal conductivity variations.

Even when tectonism has long since disappeared the case studies presented in this review have demonstrated that the memory of fluid movement may often be preserved in *mature and cratonic continental crust.* This may either be as free fluids trapped in sub-horizontal fracture zones, particularly in the lower crust, or as grain-boundary graphite precipitated after peak metamorphism. Graphite is also an important source of conductivity in the form of metasediments trapped in *continent-continent palaeosuture zones.* Given the substantial disruption in structure and composition during the evolution of continental crust, these conductive structures can provide important marker horizons to unravel its complex tectonic history. Multi-disciplinary research programs in specific regions of stable continental crust will be needed to aid our understanding of the sources of enhanced electrical conductivity. Some important questions requiring clarification are as follows. What are typical geometries of lower crustal high conductivity layers? Are their vertical conductivities substantially different from their horizontal conductivities? Can we distinguish between different sources of high conductivity (fluids, partial melt, graphite, etc)? Can we determine the origin of these sources? What are the necessary conditions for their existence over geological time scales? What do they tell us about the evolution of continental crust?

In addition to fluids mobilised during tectonism, there is now sufficient evidence to demonstrate the existence of meteoric water circulation to great depths in the continental crust (e.g. Meissner & Wever, 1992 and references therein). In some *extensional or strike-slip dominated terrains*, circulation may extend to at least

20 km depth whilst in *palaeosuture zones* meteoric and metamorphic fluids may be guided to the lower crust and beyond. Jones *et al.* (1992b) demonstrate convincingly the ability of MT to image a major strike-slip fault whose conductivity arises from such channelling of meteoric water. It is worth noting that in this case study as in many other geological environments, steeply dipping structures in the upper crust can be well-resolved by MT if they are conductive. This is not usually the case with seismic reflection which often has difficulties with resolving sub-vertical structures. The presence of such structures exacerbates the problems involved in relating surface features, for which good geological control is usually available only to depths of $\sim$1 km, to seismic images at depths usually greater than $\sim$10 km. In my view, MT data could provide crucial information in the depth range 1–10 km and the results would be of great interest to many structural and metamorphic geologists. So far, this interaction between EM practitioners and other earth scientists has been limited (e.g. EM papers rarely appear in geological journals). If such interaction can be encouraged, I suspect that EM induction studies will make substantial contributions to our understanding of second order plate tectonic processes and, in particular, the evolution of continental crust.

Acknowledgments

The paper would not have been possible without the generous response to my request for (p)reprints from colleagues in IAGA Working Group I-2. In particular, I would like to thank Alan Jones for suggesting, and the Working Group agreeing, that I present this review at the 11th Workshop on EM Induction in the Earth. He and Rosemary Hutton also provided continual practical help until the paper's completion. IAGA Working Group I-2, University College Galway and the Applied Geophysics Unit gave welcome financial support.

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