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ARTICLE

Association of Variations in the Dynamics of the Lithosphere with Sea Temperature

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ABSTRACT

Variations in the dynamics of the oceanic lithosphere are important at the societal and research levels because geological activities are associated with these variations. At any given section of the lithosphere, the time in which its typical geophysical parameters vary is considerably smaller than section's age. The lithosphere can, therefore, be assumed to proceed from one state of dynamic equilibrium to another displaced deferentially. When these conditions are accounted for in the thermal analysis of the oceanic lithosphere, the earth's internal heat flux through the lithosphere is found to be an adiabatic invariant. Lithosphere physical parameters exhibit constant change and linearity. These findings simplify analysis of heat and work interactions between oceanic lithosphere and continents, lithosphere dynamics, and deep mantle heat transfer. The temperature of the solid earth remains unchanged for the foreseeable future, and variations in sea temperature vary the intensity of geological activities. If sea temperature increases, the geological activities increase and vice versa. Relevant equations are derived using this thermal analysis of the lithosphere and validated based on observations and the work of others. In addition, the analysis reveals that the eleven-year solar constant cycle is capable of inducing 1.56×10^{16} J yr^{-1} of geological activities.

1. Introduction

Publication[1] associates variations in the observed geological activities with geomagnetic activity resulting from the eleven-year solar constant cycle. Because sea temperature varies with the cycle as well, variations in sea temperature should not be excluded from consideration. Sea temperature variations may well be the cause of the observed variations in geological activities. Reference [2] summarizes the controversy relative to the observed pattern of abyssal hills around the axis of mid-ocean ridges (MOR) and includes references for further reading on the subject. Milankovitch cycle and climate change are considered as the likely causes of the observed variation in the texture of the lithosphere around MOR. Paper [3] explores possible link between climate change and tectonics, and [4] derives mathematical relationships between variations in sea temperature and energy of plate tectonics.

Climates generally alter sea temperature, and in this work, the correlation between dynamics of the lithosphere and sea temperature will be demonstrated theoretically based on the thermal analysis of the lithosphere. Refer-

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ence [5] presents a simplified solution of the Fourier equation for the lithosphere. Textbooks [6,7] solved the Fourier equation using a more mathematically complex approach. However, the solutions are for a steady state where all of the variables including ocean temperature do not vary with time. These references address the effect of climatic change on the continental crust's geothermal gradients. They do not address the impact of climate change on the geothermal gradients of the oceanic lithosphere.

Sea temperature appears to be variable with climate change. The study [8] reported variation in the temperature of the abyssal brine of the southern oceans, and these oceans make up a substantial portion of the hydrosphere. In addition they include large segments of the plate tectonic system. The observed abyssal warming could have an effect on the thermodynamics of tectonics and should be accounted for in the thermal analysis of the oceanic lithosphere. Most of the oceanic lithosphere is thick, and the observed small variation in sea temperature is generally presumed to have negligible impact on the thermal structure or dynamics of the lithosphere. However, at midocean ridges the lithosphere is thin or yet to be formed. The observed small variation in sea temperature thus cannot be ignored; its effect is amplified by the large surface area of the lithosphere at the ridges. If sea temperature variation can alter the thermodynamics of midocean ridges, then the entire dynamics of the lithosphere of the earth can be impacted. Midocean ridges are the drivers of the oceanic lithosphere after all.

Reference [9] indicates that the new oceanic lithosphere is generated at midocean ridges at the rate of 2-20 cm yr^{-1}, which is typically referred to as annual spreading of ocean floor. It takes about 200 million years to renew the entire oceanic floor. For the foreseeable future, average ocean floor spreading is infinitesimal compared with the total width or length of the oceanic lithosphere. Consequently, lithosphere transformations inherently assume states of dynamic equilibrium displaced deferentially apart. Therefore, the Fourier equation is re-analyzed in this manuscript to account for variation in sea temperature and lithosphere dynamic equilibrium. The mathematical conclusions are interesting and differ from the current understanding: the earth's internal heat flux through the lithosphere is constant and changes in lithosphere dynamics are associated with sea temperature variations, which are in turn correlated with the climates. Other findings include constant change of lithosphere spreading and thickness and constant temperature of the solid earth. The objective of this work is to demonstrate these conclusions mathematically and validate the derived equations based on observations and the work of others. This has merits at the societal and scientific levels for it enhances our understanding of the relationship between lithosphere dynamics, geological activities, and climates.

2. Mathematical Model

2.1 Expansion of Geophysical Parameters in Series

As discussed in the introduction section, the motion of the oceanic lithosphere and motion variation occur infinitesimally with time. Consequently, the geophysical parameters of the lithosphere vary infinitesimally as well. The parameters vary continuously and they can be considered as continuous functions. Also, they can be assumed to be differentiable, and the geophysical parameters may be expanded in a Taylor Series. An arbitrary parameter of the lithosphere, Ψ(t), can be expanded as follows:

$$\Psi(t)=\Psi(t_0)+[d\Psi(t)/dt]_{t=t0}\,dt+R, \quad (1)$$ [10]

Where the zero prefix stands for an initial or a reference period of time and, dt, is an infinitesimal period of the time, t. The remainder, R, includes infinitesimal dt^2 and higher order terms and can be neglected. Equation (1) yields $d\Psi(t)/dt=[d\Psi(t)/dt]_{t=t0}$. The derivative $[d\Psi(t)/dt]_{t=t0}$ is equal to the slope of the function, Ψ(t), calculated at $t=t_0$; it is a constant number. All of the geophysical parameters can thus be considered to vary linearly in the proximity of an initial state. This mathematical approach simplifies the solution of the Fourier equation with variation in the temperature of ocean floor as will be demonstrated in the succeeding paragraphs.

2.2 Dynamic Equilibrium

For practical purposes, the physical parameters associated with the dynamics of the oceanic lithosphere have variation time, t, that is considerably smaller than lithosphere age at virtually every section. Lithosphere thermodynamic transformations are therefore displaced differentially from each other, and they may be considered in dynamic equilibrium. Referring to Fig. 1, if the lithosphere assumes states of dynamic equilibrium at the time, t_0, and, t, displaced differentially, then $F_0\ v_0$=constant and F v=constant. However, $F_0\ v_0 \neq F\ v$ and

$$F\ v-F_0\ v_0=\Delta E=\Delta M\ L_f \quad (2)$$

Where

F =Force per unit length of MOR between oceanic lithosphere and surroundings, N m^{-1}.

v =Spreading of the oceanic lithosphere, m yr^{-1}.

E =Energy exchanged with the surroundings per unit length of MOR, J yr^{-1} m^{-1}.

M=Average mass of the re-generated new basalt per unit length of MOR, kg $yr^{-1}m^{-1}$.

L_f=Basalt latent heat of solidification calculated at the pressure of the deep mantle, J kg^{-1}.

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The last term of Eq. (2) is obtained from [4]. On the other hand, $\Delta M=\rho(L\, v-L_0\, v_0)$, where L is the height of the regenerated basalt measured in meters and ρ is the density of basalt in kg m^{-3}. Because, L, and, v, vary deferentially, $\Delta M=\rho[(L_0+\Delta L)\,(v_0+\Delta v)-L_0\, v_0]=\rho(L_0\,\Delta v+\Delta L\, v_0+\Delta L\,\Delta v)$. Variation in v and L are too small compared with their values and $L_0\,\Delta v+\Delta L\, v_0\approx\Delta(L\, v)\approx\Delta(L_0\, v_0)$. Since L, v, L_0, v_0 are related to states of dynamic equilibrium, they are constants and $\Delta(L\, v)=\Delta(L_0\, v_0)=0$. The term $L_0\,\Delta v+\Delta L\, v_0$ is thus negligible and $\Delta M\approx\rho\,\Delta v\,\Delta L=\rho\,\Delta h\,\Delta v$. Where h is the height of midocean ridges above sea floor. Accordingly, Eq. (2) yields

$$\Delta v=\Delta E/(\rho\, L_f\,\Delta h)=k_1 \quad (3)$$

Because the ratio E/h is proportional to the total energy produced by tectonics to that of the potential energy of midocean ridges, which is a constant, the ratio ΔE/Δh is a constant as well. The right hand side of Eq. (3) is therefore equal to the constant k_1. Or, variation in lithosphere spreading, Δv, is the same for any two consecutive states of dynamic equilibrium.

Referring to Fig. 2, an arbitrary section of the lithosphere at a distance, x, from midocean ridges has an age $G=2\,x/v$ and a thickness $d=k_2\,G^{1/2}$, where k_2 is a constant of proportionality. For an initial ocean floor spreading v_0 at an initial time t_0, the lithosphere section at the distance x_0 from midocean ridges has an age $G_0=2\,x_0/v_0$ and a thickness $d_0=k_2\,G_0^{1/2}$. For an infinitesimal variation in ocean floor spreading from, v_0, to, v, in an infinitesimal period of time $dt=t-t_0$ and neglecting the remainder R, Eq. (1) for, d(v), and, v, instead of, Ψ, and, t, respectively yields

$$[d(v)-d(v_0)]/\Delta v=\Delta d/\Delta v\approx[(1/2)\,k_2\,G^{-1/2}\,dG/dv]_{v=v0}=k_3 \quad (4)$$

Where

G =Section age for floor spreading v, yr.

v =Ocean floor spreading, m yr^{-1}.

d(v) =Lithosphere thickness at a distance, x_0, from midocean ridges, m.

Because the time of lithosphere dynamic variation, t, is considerably smaller than the age of the lithosphere at the section in consideration, then $G\approx G_0$ and the term $[(1/2)\,k_2\,G^{-1/2}\,dG/dv]_{v=v0}$ of Eq. (4) is equal to a constant k_3. Equation (3) reveals that variation in ocean floor spreading, Δv, is a constant that has the same value between any two consecutive states of dynamic equilibrium. Equation (4) on the other hand indicates that variation in the thickness of the lithosphere, Δd, is another constant having the same value between any two consecutive states of dynamic equilibrium. These are intrinsic characteristics of the dynamics of the lithosphere resulting from the nature of magmatic process and large ages of typical sections of the lithosphere.

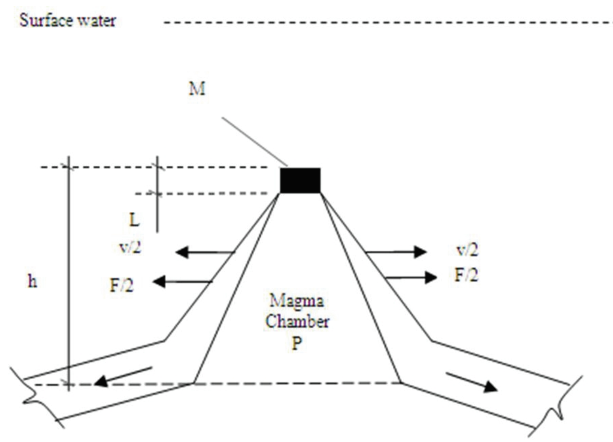

Figure 1. A schematic representation of midocean ridges and lithosphere plates based on [9], not to scale. The dark area represents the regenerated new mass M of basalt per unit length of midocean ridges. F is the force imparted by the lithosphere on the surroundings, v is ocean floor spreading, and P is the magma pressure. When magma pressure varies, the height, L, of the basalt produced varies as well.

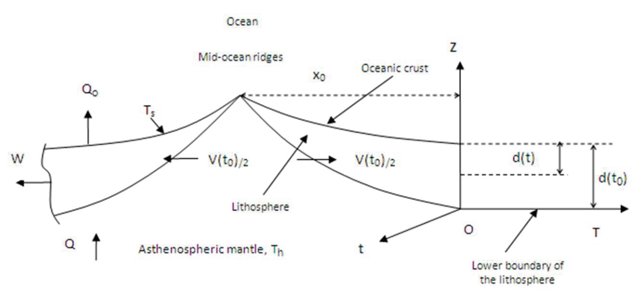

Figure 2. A section of the oceanic lithosphere per unit length of midocean ridges, located at a distance x_0 from the ridges based on [9], not to scale. The mobile coordinate system (t, T, z) moves with the lower boundary of the solid rocks of the lithosphere. The coordinates are time, t; temperature, T; and position, z. The position of any point of the lithosphere is measured from the lower boundary of the lithosphere. The location of this boundary is variable with lithosphere dynamics and the origin of the coordinate system, O, translate up for an increase in ocean floor spreading.

2.3 Re-analysis of the Fourier Equation

In Fig. 2, a section of the lithosphere along with a Cartesian coordinate system is presented based on [9]. The section is located at a distance x_0 from midocean ridges and has an initial thickness of $d(t_0)$. The lithosphere thickness is variable with time. The Fourier equation applies only to the solid rocks of the lithosphere. Accordingly, a mobile Cartesian coordinate system is considered for time,

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temperature, and vertical position (t, T, z) having origin, O, at the lower boundary of the lithosphere above which the lithosphere is assumed to be made of solid rocks. The lower boundary of the lithosphere is the interface location between solid rocks of the lithosphere and the ductile asthenospheric mantle. Below this interface, the ductile mantle exists and the Fourier equation does not apply. As the thickness of the lithosphere varies with ocean floor spreading, the position of the interface boundary moves up or down. The mobile coordinate system thus translates in the, z, direction with the lower boundary of the solid lithosphere. For oceanic lithosphere having increasing dynamics, Fig. (2) presents the related scenario. The thickness of the lithosphere, d, decreases infinitesimally from $d_0=d(t_0)$ to d=d(t) according to the equality $d=k_4/v^{1/2}$, where k_4 is a constant of proportionality. In the region $d(t_0)$-d(t), the Fourier equation does not apply for it is assumed to convert to a ductile mantle. The Fourier equation for an arbitrary rectangular parallelepiped finite element of the solid lithosphere having a unit volume and a height that is equal to, dz, follows:

$$dT(t, z)/dt=[k/(\rho\ C_p)]\ \partial^2T(t, z)/\partial z^2+Q_g/(\rho\ C_p)\ ,\quad (5)$$ [11]

Where T(t, z) is the average temperature of the finite element, °K; t, is the time in seconds; k, is the thermal conductivity of the selected arbitrary finite element of the solid lithosphere, J s^{-1} m^{-1} $°K^{-1}$; ρ is the density of the rocks of the finite element, kg m^{-3}; C_p, is the specific heat of the finite element, J kg^{-1} $°K^{-1}$; and z is the vertical location of the finite element measured from the interface between the solid rocks of the lithosphere and the asthenospheric mantle. This interface moves upward because the thickness of the solid rocks decreases with an increase in ocean floor spreading. The finite element selected is small enough to assume that its thermodynamic properties to be reasonably uniform. The heat generated per unit volume, Q_g, is equal to zero. The partial derivative of the temperature, ∂T(t, z)/∂z, is the slope of the temperature function T(t, z) on the T-z plane. It will be indicated by the symbol, S, for slope

$$S(t,z)=\partial T(t,z)/\partial z \quad (6)$$

Because variations occur infinitesimally and continuously with time, the slope function such defined, S(t, z), may be expanded in a Taylor Series around an arbitrary state of equilibrium. Neglecting the remainder R

$$dS(t, z)=[\partial S(t, z)/\partial t]_{t=t0}\ dt+[\partial S(t, z)/\partial z]_{z=z0}\ dz \quad (7)$$

$$dS(t, z)/dt=[\partial S(t, z)/\partial t]_{t=t0}+[\partial S(t, z)/\partial z]_{z=z0}\ dz/dt \quad (8)$$

The position, z, of any point of the solid rocks of the lithosphere measured from the origin, O, varies with respect to the mobile coordinate system; it is unchanged with respect to a stationary coordinate system. The stationary coordinate system will be defined by (t, T, ζ) having origin Ω that coincides with the initial position, O, of the mobile coordinate system. Where ζ is the position of the finite element in consideration with respect to the stationary coordinate system. The relationship between positions of the two coordinate systems follows:

$$z=\zeta-\Delta d \quad (9)$$

Where, z, is the position of the finite element with respect to the mobile coordinate system, and ζ is the position of the same finite element with respect to the stationary coordinate system. For $v=v_0$ at the time $t=t_0$, the thickness of the lithosphere is equal to $d(t_0)$. After an infinitesimal period of time dt, v=v(t) and the thickness of the lithosphere is equal to d(t), where $t=t_0+dt$. The function d(t) thus exists and can be assumed continuous and differentiable between t_0 and t. The difference between the initial thickness and the final thickness of the lithosphere $d(t_0)$-d(t) is equal to Δd, Fig 2. Neglecting R, Eq. (1) yields

$$\Delta d=d(t_0)-d(t)=-[d\{d(t)\}/dt]_{t=t0}\ dt \quad (10)$$

On the other hand, Eq. (4) of the dynamic equilibrium section yields

$$d(v_0)-d(v)=\Delta d=-k_3\ \Delta v=m_1 \quad (11)$$

Where m_1 is a constant whose value is the same between any two consecutive states of dynamic equilibrium. The infinitesimal time, dt, can be obtained from Eq. (10) and Eq. (11)

$$dt=-m_1/[d\{d(t)\}/dt]_{t=t0}=m_2 \quad (12)$$

Where m_2 is a constant for the slope of the d(t) function, $[d\{d(t)\}/dt]_{t=t0}$ is a constant number. This slope calculated at $t=t_0$ is about the same for all of the states of dynamic equilibrium, as this is clear from Eq. (4). If this equation is reproduced for d(t) and, t, instead of d(v) and , v, and for section age nearly unchanged, $G \approx G_0$, the initial slope $[d\{d(t)\}/dt]_{t=t0}$ is equal to a constant whose value is equal to (1/2) $k_2\ G_0^{-1/2}\ [dG/dt]_{t=t0}$. The value of this last formula is the same, constant, for any arbitrary initial state of dynamic equilibrium.

Equation (12) indicates that the states of dynamic equilibrium are equally displaced in time, they reach dynamic equilibrium in equal periods of time. For the equilibrium state at the initial conditions, the thickness $d(t)=d(t_0)$ and Δd=0.This yields to z=ζ based on Eq. (9) and dz/dt=dζ/dt. The position of the finite element, ζ, with respect to the stationary coordinates is unchanged with time and dζ/dt=0. As a result, dz/dt=0 at the initial conditions and Eq. (8) yields

$$dS(t, z)/dt=[\partial S(t, z)/\partial t]_{t=t0} \quad (13)$$

The right hand of Eq. (13) includes a partial derivative with respect to the time only, ∂S(t, z)/∂t. Or the depth, z, can be treated as a constant. Accordingly, $\partial S(t, z)/\partial t=dS(t, z)/dt=[dS(t, z)/dt]_{t=t0}$=A=constant based on series expansion of Eq. (1) and for R=0. This last equality yields to dS(t, z)/dt=A and its integration yields

$S(t, z)=A\ t+C$ (14)

Where C is a constant of integration. Eq. (14) reveals that the slope function, S(t, z), is independent of z. Therefore, $\partial S(t, z)/\partial z=0$, and based on Eq. (6), $\partial^2 T(t, z)/\partial z^2=0$. Because there is no heat production in the lithosphere, $Q_g=0$, the Fourier Eq. (5) simplifies

$dT(t, z)/dt=0$ (15)

This equation (15) indicates that the temperature profile, T(t, z), of the oceanic lithosphere is time independent and, z, is the only variable. As a result, $\partial T(t, z)/\partial z=dT(t, z)/dz$. Per definition of S(t, z), Eq. (6), then $\partial T(t, z)/\partial z=S(t, z)=dT(t, z)/dz$. Based on this last equality and Eq. (14)

$dT(t, z)/dz=A\ t+C$ (16)

Equation (16) can be used to derive the temperature profile of the lithosphere as it displaces from one state of dynamic equilibrium to another. The time, t, of Eq. (16) cannot exceed the differential time, dt, between two consecutive states of dynamic equilibrium. For a complete transition from one state to another, t, must be equal to, dt, and

$dT(t, z)/dz=A\ dt+C$ (17)

As discussed earlier, dt is a constant, Eq. (12), and Eq. (17) yields to

$dT(t, z)/dz=A\ m_2+C=m_3$ (18)

Where m_3 is a constant whose value is equal to $-Q/(k\ A_s)$ based on the solution of equations (5), (15), (18), the equality $\partial T(t, z)/\partial z=dT(t, z)/dz$ obtained earlier, and for $Q_g=0$. The solution is a straightforward one, available in typical heat transfer text books, example [11]. Where Q is the earth's internal heat flux through the oceanic lithosphere, J s^{-1}, and A_s is a constant that is equal to the surface area of the sphere enclosing the lithosphere at the finite element in consideration, m^2. The sphere has a center that coincides with the center of the earth. The thermal conductivity, k, of every finite element is reasonably uniform and constant but varies in value with depth from one finite element to another. Consequently

$Q=-k\ A_s\ m_3=-k\ A_s\ dT(t, z)/dz=\text{constant}$ (19)

Solution of Eq. (18) for a state infinitesimally displaced from an initial state of equilibrium follows:

$T(t, z)=m_3\,(z-z_0)+T(t_0, z_0)$ (20)

$T(t, z)=dT(t, z)/dz\ x\ (z-z_0)+T(t_0, z_0)$ (21)

Where, t, and, z, are time and position infinitesimally displaced from the initial equilibrium conditions t_0 and z_0. Equation (21) gives the temperature of the first state of dynamic equilibrium displaced infinitesimally from the initial state of equilibrium. The procedure can be repeated for the second state of dynamic equilibrium by considering the first state as an initial condition until all of the period of time in consideration is simulated.

Equation (21) can be transformed using the stationary coordinate system (t, T, ζ)

$T(t, \zeta)=dT(t, \zeta)/d\zeta\ x\ (\zeta-\Delta d-\zeta_0)+T(t_0, \zeta_0)$ (22)

For this coordinate transformation, $z=\zeta-\Delta d$ from Eq. (9). Because Δd is a constant based on Eq. (11), then $d\zeta/dz=1$. As a result, coordinate transformation of Eq. (18) yields to an equality between dT(t, z)/dz and dT(t, ζ)/dζ. The positions ζ and ζ_0 are equal because they define the position of the same finite element at the time $t=t_0$ and t=t with respect to the stationary coordinate system. Therefore $\zeta-\zeta_0$ of Eq. (22) is equal to zero. This discussion and equations (15), (19), and (22) yield respectively to the following summary:

$T(t, z)=T(t_0, z_0)=T(t, \zeta)=\text{constant}$ (23)

$Q=-k\ A_s\ x\ dT(t, \zeta)/d\zeta=\text{constant}$ (24)

$T(t, \zeta)=-dT(t, \zeta)/d\zeta\ x\ \Delta d+T(t_0, \zeta_0)$ (25)

3. Results

Equation (23) represents an infinite number of planes parallel to the t-ζ coordinate plane, and Eq. (25) is an infinite number of inclined planes. The intersection between these two planes defines the temperature profile that satisfies lithosphere thermodynamics, which is a surface in the three dimensional space. The projection of this surface on any plane parallel to the T-ζ coordinate plane is a curve presented in Fig. 3 and Fig. 4. Boundary conditions are the temperature of oceanic crust, T_s, and asthenospheric mantle temperature T_h.

At the initial time $t=t_0$, Δd=0 and based on Eq. (25), $T(t, \zeta)=T(t_0, \zeta_0)$ for every point of the curve, Fig. 3. With an increase in the dynamics of the lithosphere, the temperature profile changes for Δd≠0. If ocean floor spreading increases, then Δd is positive and the quantity $-dT(t, \zeta)/d\zeta\ x\ \Delta d$ is positive for the geothermal gradients have a negative signs, which gives $T(t, \zeta)>T(t_0, \zeta_0)$ based on Eq. (25). The temperature curve translates vertically as illustrated in Fig. 4. Or, the finite element assume higher temperature but its temperature remains on the temperature profile at which location the earth's internal heat flux to the oceanic lithosphere, Q, is constant. Any point on the temperature curve observes the same constant earth's internal heat flux, Q, based on Eq. (24). When the temperature of ocean floor increases by dT_s, the floor assumes the value of the isotherms just below it such that the temperature of the oceanic floor remains on the temperature profile T(t, ζ). The earth's internal heat flux through the oceanic crust, Q, thus remains unchanged and

$dQ/dT_s=0$ (26)

Where T_s is sea temperature at ocean floor. The mechanical work delivered by the oceanic lithosphere to the surroundings, W, can be calculated by the energy balance of the lithosphere. Based on Fig. 2

$M_L\ C_p\ dT(t, z)/dt=Q-Q_o-W$ (27)

Where

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M_L = Mass of the oceanic lithosphere, kg.

C_p = Average specific heat of the lithosphere, J kg^{-1} °K^{-1}.

T(t, z) = Temperature of the lithosphere, °K.

W = Mechanical work of the lithosphere exchanged with its surroundings, J yr^{-1}.

Q = Earth's internal heat flux to the oceanic lithosphere, J yr^{-1}.

Q_o = Earth's internal heat flux rejected to the ocean, J yr^{-1}.

Based on Eq. (15), dT(t, z)/dt=0 and Eq. (27) yields $W=Q-Q_o$. Derivation of both sides of this last equality with respect to the sea temperature at ocean floor, T_s, bearing in mind that $dQ/dT_s=0$, Eq. (26), gives

$$dW/dT_s = -dQ_o/dT_s \quad (28)$$

Equation (28) associates variation in lithosphere dynamics, W, or geological activities, with sea temperature.

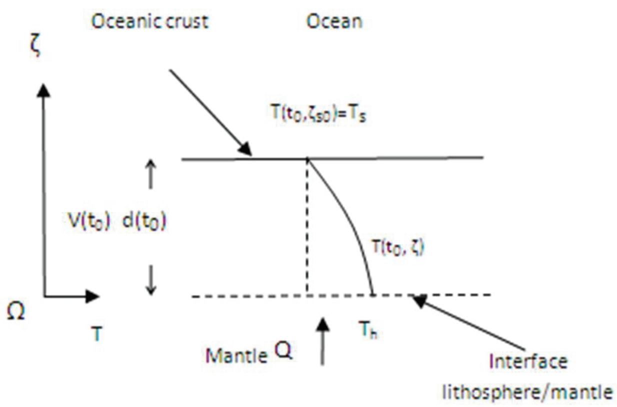

Figure 3. Projection of the temperature profile of the oceanic lithosphere on the T-ζ plane of the stationary coordinate system for the initial state of dynamic equilibrium when the time $t=t_0$ and ocean floor temperature is equal to T_s. Where ζ is the vertical position of an arbitrary finite element of the lithosphere measured from the initial location of the interface between solid rocks of the lithosphere and the ductile asthenospheric mantle having temperature T_h .

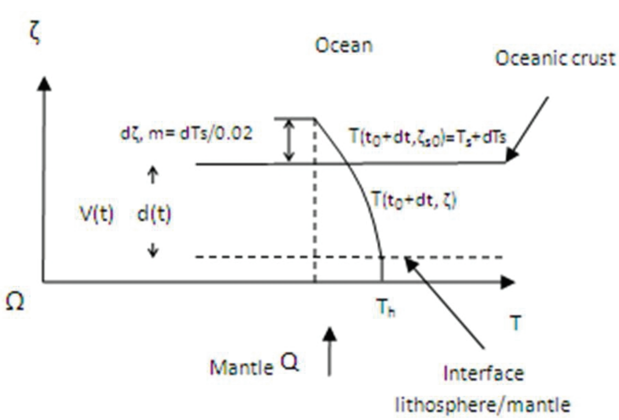

Figure 4. The temperature profile projection of Fig. 3 after the temperature of the ocean floor, T_s, has increased infinitesimally by dT_s in the infinitesimal period of time dt.

4. Application Equations

In this section, Eq. (28) will be used as basis to develop practical equations to validate the theoretical thermal analysis of the lithosphere. Also, as an application example, the geological activities resulting from the eleven-year solar constant cycle will be calculated. The heat flux to ocean Q_o may be estimated

$$Q_o = U A (T_h-T_s) \quad (29)$$

Where

U = Overall heat transfer coefficient between asthenospheric mantle and ocean, J yr^{-1} m^{-2} °K^{-1}.

A = Average heat transfer area of the oceanic lithosphere, m^2.

T_h = Temperature of the asthenospheric mantle, 1553.20 °K.

T_s = Temperature of surface water at ocean floor, 274.20 °K.

The value of the earth's internal heat to the ocean Q_o is equal to 70% of the total heat produced in the earth's core of 1.5 x 10^{21} J yr^{-1}, [12]. Asthenospheric mantle and sea temperatures are obtained from [9]. Assuming that the overall heat transfer coefficient remains unchanged with the observed small rise in sea temperature, equations (28) and (29) yield

$$dQo= -Qo\ dTs/(Th-Ts) \quad (30)$$

$$dQ_o/dt = -8.21 \times 10^{17} dT_s/dt \quad (31)$$

$$dW/dt= 8.21 \times 10^{17} dT_s/dt \quad (32)$$

Equation (32) calculates annual trend of geological activities by knowing total variation in the temperature of ocean floor. This temperature variation must be equal to variation in sea surface temperature. Otherwise, the thermohaline circulation would cease frequently which is not observed. In order to correlate trends of geological activities with the eleven-year solar constant cycle, empirical equation correlating variation in the solar constant and sea surface temperature is required. Accordingly, Table 1 is prepared based on observed seasonal variation. Using thermodynamics, it can be demonstrated that

$$\Delta T_s/T = -\alpha\ \Delta I/I \quad (33)$$

Where

ΔT_s=Variation in sea surface temperature, which is equal to variation in the temperature of ocean floor, °K.

T = Sea surface temperature, 286.90 °K.

α = Constant of proportionality, dimensionless.

I = Average value of the seasonal solar constant, 1367 W m^{-2}.

From Table 1, the average value of α is 0.0378, and Eq. (33) simplifies

$$\Delta T_s = -0.008\ \Delta I \quad (34)$$

5. Discussion and Conclusions

It is well known that slowly-occurring processes with time

Table 1. Observed monthly average Sea Surface Mean Temperature, T, and solar constant, I. The observed temperature is obtained from [15] for the Base Period 1901 to 2000. Values of seasonal solar constant are provided by [16].

Month/Description	Solar Constant	Observed	Observed	α
	I, W/m^2	T, °C	T, °K	
January	1409.10	15.80	289.00	0.0390
February	1399.00	15.90	289.10	0.0387
March	1379.50	15.90	289.10	0.0382
April	1358.90	16.00	289.20	0.0376
May	1352.00	16.30	289.50	0.0374
June	1323.80	16.40	289.60	0.0366
July	1321.50	16.40	289.60	0.0365
August	1330.70	16.40	289.60	0.0368
September	1350.80	16.20	289.40	0.0373
October	1372.50	15.90	289.10	0.0380
November	1396.30	15.80	289.00	0.0387
December	1408.10	15.70	288.90	0.0390

are characterized by linearity and conservation of processes characteristic parameters. Variations in the forces between the planets of the solar system occur slowly, which yield to conservation of the angular momentum of planets and length of the semi-major axis of the orbital planes. When the length of the string of a pendulum is gradually and slowly varied with time, the ratio between pendulum energy and frequency remains constant, commonly indicated as adiabatic invariant ratio. Climate change parameters vary deferentially with time and linearity of climate parameters is observed where natural variability is small or can be filtered. Reference [13] reveals that stratospheric cooling occurs at levels following climate perturbations. At each level temperature trend is practically linear. So is sea level rise [14]. The rise appears to proceed at campaigns, each having practically linear trend. The same must be true for the lithosphere. The spreading of ocean floor is so infinitesimal compared with floor's length and its variation is infinitesimal as well. The physical parameters of the oceanic lithosphere associated with this spreading must exhibit linearity, constant change, or adiabatic invariant values. Based on this work, spreading of the oceanic lithosphere varies linearly with time, and its rate of change remains constant with changes in ocean floor spreading. The same is true for the thickness of the oceanic lithosphere at virtually every section. These conclusions are intrinsic properties of the oceanic lithosphere resulting from the nature of magma generation and solidification at midocean ridges. The process is so slow compared with the time required to regenerate the full ocean floor.

As a result, Fourier equation re-analysis yielded to a constant earth's internal heat flux through the oceanic lithosphere. This adiabatic invariant nature of the heat flux is an intrinsic property of the solid earth. Any variation in the temperature of ocean floor adhering to the oceanic lithosphere does not alter the value of the heat flux. The temperature profile of the lithosphere remains unchanged; it only translates up and down with variations in sea temperature and the earth's internal heat flux in every section remains the same. While the temperature of each section of the solid rocks of the oceanic lithosphere varies, the thickness of the rocks varies as well, proportionally, to maintain a constant heat flux through the oceanic lithosphere.

If the sphere of the solid earth enclosing the interface between lithosphere and asthenospheric mantle is considered as a thermodynamic system. The system as defined only exchanges earth internal heat with the oceanic and continental lithospheres located above it. Matter exchanged is replaced in kind and system mass may be assumed as a constant. Therefore, the system as defined may be considered as a closed thermodynamic system. Because, the generated heat rate in the earth's core is constant for the foreseeable future and the internal heat flux to the oceanic lithosphere is adiabatic invariant, then the internal heat flux to the continental lithosphere is constant. The mass and heat balance of the thermodynamic system as defined leads to a constant temperature of the solid earth.

A constant heat flux in spite of variation in sea tem-

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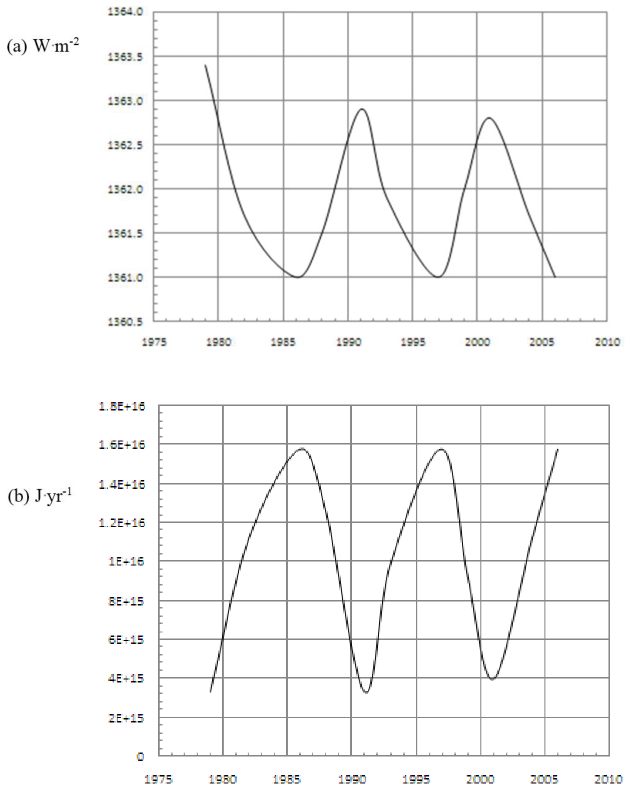

Figure 5. a) Eleven-year solar constant cycle obtained from [17]. b) Calculated annual trend of geological activities resulting from the eleven-year solar constant cycle only.

perature yields to Eq. (28), which is an application of the first law of thermodynamics on the oceanic lithosphere. Variation in sea temperature varies the mechanical energy exchanged between the oceanic lithosphere and the surrounding continents. An increase in sea temperature with time decreases the heat flux from the lithosphere to the ocean, Q_o. The term, dQ_o/dT_s, has a negative sign and the geological activities increases. Conversely, if ocean temperature decreases, the heat flux from the lithosphere to the ocean increases and the term dQ_o/dT_s is positive. The geological activities decrease as a result.

It should be noted that variation in the heat transferred between the mature lithosphere and ocean with the observed sea temperature rise is negligible for this lithosphere is thick. The lithosphere at midocean ridges on the other hand is thin or yet to be formed, and variation in the heat transferred is not negligible. Therefore, midocean ridges experience thermodynamic changes with sea temperature variation, and these changes are transmitted to the entire lithosphere. Every section of the mature lithosphere will thus experience constant change and linearity of ocean floor spreading and section thickness. Equations (23), (24), (25), (26), and (28) are thus valid for the entire oceanic lithosphere including mature sections as well.

Causes of ocean temperature variation include but not limited to climate change, seasonal variation, and eleven-year solar constant cycle. Contrary to common perception, surface temperature increases when the value of the solar constant decreases and vice versa. Average sea surface temperature is on the rise and it assumes maximum values in the northern hemispheric summer when the value of the solar constant approaches minimum values. These periods of solar constant minima should observe higher intensity of geological activities than average. They include but not limited to seismic activities, volcanic eruptions, and rise of midocean ridges. For the observed present sea temperature rise of 0.6 °K, Eq. (32) gives annual trend of geological activities of 4.93×10^{17} J yr^{-1}. The calculated value by [4] using thermodynamics is 1.05×10^{17} J yr^{-1}, and the observed trend is 3.0×10^{17} J yr^{-1}. The theoretical thermal analysis, therefore, yields results that are of the same order of magnitude of observations and work of others. Accordingly, annual trend of geological activities resulting from the eleven-year solar constant cycle is calculated using equations (32) and (34) and presented in Fig. 5. The figure reveals that at solar constant minima, the eleven-year solar constant cycle is capable of inducing 1.56×10^{16} J yr^{-1} of geological activities, equivalent to magnitude 7.6 earthquakes. The observations of Gulyaeva [1] are in agreement with this work.

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