DETERMINISTIC MODEL FOR PREDICTING TEMPERATURE CHANGE FROM GREEN HOUSE GASES

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CERTIFICATION

The thesis titled: DETERMINISTIC MODEL FOR PREDICTING TEMPERATURE CHANGE FROM GREEN HOUSE GASES by: NDACEKO USMAN IBRAHIM Registration Number: M.Eng/SEET/2008/1956 meets the regulation governing the award of Masters of Engineering of Federal University of Technology, Minna, Niger State and it is approved for its contribution to the scientific knowledge and literary presentation

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6/3/13

DECLARATION

I hereby declare that this thesis titled "Deterministic model for predicting temperature change from greenhouse gases".

Is a collection of my original research work and it has not been presented for any other qualification anywhere. Information from other sources (published or unpublished have been duly acknowledged.

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ABSTRACT

A deterministic model for predicting temperature change from green house gases was developed and simulation of the model equation was done by inputting the parameters such as specific heat capacity of the gases, density of the gases, heat transfer coefficient of the gases, minimum temperature change of the materials to radiate heat, Stefan Bolt man's Constant, emissivity of the materials at varying time. The model was simulated using Mat Lab. Experiment to demonstrate green house effect. Materials used are charcoal, chippings, laterite, organic soil and sharp sand. In this research work the developed model was tested using data generated from green house effect. The experimental and simulated results were obtained and the graphs were plotted to show the relationship and the trend of the results. The slopes of the experimental graph are 0.18000, 0.15733, 0.126667, 0.153333, and 0.14000. The average slope of experimental graph is 0.151466, and the average slope of the simulated graph is 0.1511084 while the difference between the average simulated slope and the experimental one is 0.03576 percent this shows the extent of their closeness. The slope of the graphs also show that the deterministic model can serve as tool for analysis of temperature change from green house gases with respect to time.

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CHAPTER ONE

1.0 INTRODUCTION

The increases in human population and activities of man on earth have also increased correspondingly, the emission of gases to the atmosphere. Some of these gases that are emitted in the atmosphere are called green house gases. The green house gases from industrial areas are Hydrofluro carbon (HFC), parfluoro carbon (PFC), sulphur fluoride (SF) and chlorofluro carbon (CFC). These gases did not exist before 20th century (Butler, 1999). The technological development in areas of transportation and manufacturing using heavy duty engines have many positive benefits but its by - products could cause global warming or long time climate change. It is important to understand the effect of these gases on the atmosphere (Butler, 1999).

The atmosphere is made up of air and water vapor. Air consists mainly of oxygen, nitrogen, carbon dioxide, argon and other gaseous components in the form of monatomic gases with unequal charge distribution and poly atomic gases Butler, (1999). Human activities such as cooking, bush burning, combustion engines always result in some by-products e.g. carbon monoxide, nitrogen oxides, sulphur oxide, carbon floro-carbon(CFC),parfluro carbon (PFC),hydroflurocarbon (HFC) being emitted into the atmosphere. Presently the high concentrations of these green house gases have resulted in global warming on climate and change in ecosystem (Butler, 1999).

The radiations from the sun are absorbed by the green house gases and are transferred to the earth. Infrared radiations from earth are reabsorbed by green house gases and thereby increase the temperature of the earth causing global warming which categorically has impacted on human health, economic activity and ecosystem by their short and long term effects. Such green house gases include carbon monoxide, nitrogen oxide, sulphur oxide, chlorofluorocarbon, sulphur fluorides, methane, ozone, hydrogen chloride and water vapour. Sources of these green house gases include municipal activities such as refuse disposal, fuel fired combustion processes, motor vehicles, domestic heating and cooking activities (Butler, 1999).

When the green house gases are emitted, they are transported by wind current from their point of release to other parts of the atmosphere. The dispersion of green house gases is controlled by metrological parameters, atmosphere stability, and wind and speed direction. Nature has provided residence time for removal of green house gases in the atmosphere. This natural process includes reaction with hydroxyl radical, destruction at oceanic surface, precipitation, scavenging and photolysis. The rate of natural removal at our present \level of development is indeed slower than the rate of emission of green house gases into the atmosphere because of their longer residence time (Eisenhauer, 1998).

Several sources of green house gases can be identified. These include cars and other combustion engines or general fossil fuel driven vehicles. These have contributed to good standard of living enjoyed by the populace but their combustion processes have resulted in the production of carbon dioxide, carbon monoxide, chlorofluoro carbon (CFC), nitrogen oxide and other green house gases (GHG). Domestic cooking and heating are largely carried out with gas and kerosene stoves. These also provide various gases as well as soot, and flying ash that are eventually discharged into the atmosphere. The amount of gases emitted by bush burning, decomposition of agricultural waste perhaps may not be significant, but it contributes to the overall contaminant concentration in the atmosphere (Harnisch, 1995).

1.2 Aim and Objectives of the Study

The main aim of this work is to develop a deterministic model for predicting temperature

change from green house gases. The aim shall be achieved by the following objectives,

(1) Develop a mathematical model using Principles of thermodynamics
(2) Use MATLAB to solve the model equation
(3) Compare the simulated results with the experimental results and draw appropriate conclusion.

1.3 Scope of the Study

This research work covers the mathematical model of green house effect, simulation of the model, performing the experiment that demonstrates green house effect, plotting the graph of the experimental and simulated results and analyzing the rate of change of temperature with respect to time.

1.4 Justification of the Study

The emission of green house gases is causing deterioration of the earth's atmosphere. So the reason of this research work is that:

1) If Green house gases is causing temperature change of the atmosphere
2) If green house effect of the atmosphere is demonstrated by experimental . Temperature change
3) If mathematical model can be used instead of the routine experimental procedure.
4) If the rate of temperature changes with respect to time in different materials can be determined.

CHAPTER TWO

2.0 LITERATURE REVIEW

2.1 History of Greenhouse Gases

In terms of the energy balance of climate system, Houghton, (2002). Noted that although the sun's light and heat easily passes through the glass and other transportation materials, heat from other source does not. The ability to generate an artificial warming of the earth's surface was demonstrated in simple green house experiment by Houghton, (2007), using a heliothermomenter (panes of glass covering a thermometer in the darkened box) to prove an early analogy to the greenhouse. effect it. The temperature of the earth's can be augmented by the emmiter position of the atmosphere, because heat in the state of light finds less resistance in penetrating the air, than in reparsing into the air when converted into nonluminons heat: Houghton, (2002) followed up on Fourier's ideas and argued that"the atmospheric striations exercises a greater absorption upon the terrestrial than on the solar ray. However, there was still no understanding of exactly what substance in the atmosphere that was responsible for this absorption (Fleming, 1998)

Houghton, (2007) identified the absorption of the thermal radiation by complex molecules (as opposed to the primary bimolecular atmospheric constituents O_2 and molecular nitrogen). It was noted that changes in the amount of any of the radioactively active constituents of the atmosphere such as (H_2O) or CO_2 could have produced all the mutations of the climate which the researches of the geologies revealed; Eisenhauer, (1998) followed up with climate prediction based on green house gases, suggesting that a 40 % increase or decrease in the atmospheric abundance of the trace gases CO_2 might trigger the glacier advances and retreats. One hundred years later, it was found that CO_2

did indeed vary by this amount between glacial and interglacial periods. However, it now appears that the initial climate change preceded the change in CO_2 but was enhanced by it (Houghton, 2007).

Eisenhauer, (1998) reported in a work solved a set of equation linking greenhouse gases and climate change showed that a doubling of atmospheric CO_2 concentration resulted in an increase in the mean global temperature of 2 °C with considerably more warming at the poles and linked increasing fossil fuel combustion with a rise in CO_2 and its greenhouse effect. In this same study, it was equally asserted that increasing atmospheric CO_2 would lead to gradual increase in the mean temperature of the colder region of the earth. (Eisenhauer, 1998)

In the 1970s, the greenhouse gases CH_4, N_2O and CFCs were widely recognized as important anthropogenic greenhouse gases. Others included and atmospheric aerosols (suspended small particles) were also being proposed as climate forcing constituents (Charlson 1990).

2.1.1 Sources of Green House Gases

Substantial and pre-Industrial abundance of CH_4 and NO_2 are found in the tiny bubbles of ancient air trapped in ice cores. Both gases have large, natural emission rates, which have varied over past climatic changes but have sustained a stable atmospheric abundance for the centuries prior to Industrial Revolution (Butler, 1999).

The industrial abundance for a second class green house gases are the synthetic HFCs, PFCs, SF6, CFC and Halons did not exist in the atmosphere before 20th century (Butler, 1999).

2.1.2 The Effect of Greenhouse Gases on Atmospheric Temperature.

The global system that regulates the earth's temperature is very complex, but many scientists believe that the increase in temperature is caused by an increase of certain gases in the atmosphere that trap energy that would otherwise escape into space. These gases, called greenhouse gases, include CO_2 CH_4, NOX. CFCs and the ozone in the lower atmospheres. To get an idea of how an increase in these gases in the atmosphere could lead to global warming, lets first look at the role that some greenhouse gases normally play in regulating the flow of energy into and out of the earths atmosphere (Eisenhauer, 1998).

There is a natural balance between the energy coming into our atmosphere from the sun and the energy escaping from the earth back into space. A simplified description of the energy balance depicts the energy that comes in as relatively short wave length UV radiation and the energy that goes out as long wave length infrared Radiation (IR) radiation. Much of the high energy radiation from the sun is absorbed in the upper atmosphere, but some of the ultraviolet radiant energy passes through the lower energy end of the ultraviolet portion of the light spectrum reaches the earths surface and warms it. As the earth cools down again it radiates energy as infrared radiation (Eisenhauer, 1998).

The green house gases absorb some of the infrared radiation released as the earth cools. When they re-emit it, some of the energy is sent back toward the earth, rather than out into the space. If the green house gasses were not there to trap some of this (IR) energy, the earth will be about 33 ° C (60 ° F) colder than it is. These gases are called greenhouse gases because their effect is like the glass panels in a green house. Radiant energy passes through the panels to enter the greenhouse, but the panels slow down the passage of the heat energy leaving the green house. On a cold day, more energy comes into the green house than goes out, so the air in the house is warmed (Eisenhauer, 1998).

Even though the actual mechanism is very different, the net effect of the greenhouse gases in the atmosphere is similar: they too tend to trap energy that would otherwise escape, keeping the temperature of the system higher than it would be otherwise (Eisenhauer, 1998).

The green house effect causes some of the ultraviolet light that comes from the sun passes through the earth atmosphere and warms the earth. As the earth cools, it emits infrared radiation. Some of this infrared radiation is absorbed by the green house gases. When the gas particles re-emit the infrared radiation, a portion of it is sent back towards the earth. Thus, the green house gases trap some of the energy that would have otherwise escaped. This trapped energy leads to rise in temperature on the earth that are higher than they would be without the green house gases (Eisenhauer, 1998).

2.1.3 The Science of Green House Gases

The concept of global warming itself is controversial because of the complexity of our global weather system and difficulty of predicting changes in it. For example, as the concentration of the green house gases in the atmosphere increases, the temperature of the earth increases, and more water evaporates from the surface of the earth. The effect of this mechanism is difficult to predict. Water vapour itself is a greenhouse gas, so an increase in the gaseous H_2O concentration in the atmosphere would trap more (IR) radiation and could lead to an even greater increase in temperature. On the other hand, the increase in the water vapour also lead to increase in cloud. These especially at low attitudes, can reflect the incoming ultraviolet radiation back into space and lead to cooling. Meanwhile, the green house effect of the water molecules in high clouds seems to outweigh the reflective effect of the clouds. The infrared Radiant energy from UV radiation reflects back into space when the cloud is high. Therefore, the complexity of the system makes it very difficult to predict whether more water in the atmosphere will lead to warming or cooling (Butler, 1999).

2.1.4 Greenhouse Gases: Warming or Cooling

Another example of the complexity of the global warming stems from a relationship between this topic and stratospheric chemistry. If the IR radiation trapped by excess green house gases in the lower atmosphere had been allowed to move into the stratosphere as the natural cycle would have allowed, the gases in the stratosphere would have absorbed some of it and been warmed. Thus the increase in the green house gases is thought to have caused a cooling of the stratosphere. This cooling may have increased the formation of ice crystals there, and provided a surface on which ozone molecule are destroyed at a greater rate than would otherwise occur. Less ozone allows more ultraviolet radiation to reach the earth surface. This can damage the plants that take CO_2 from the air and lead to an increase in the CO_2 levels in the atmosphere. The extra CO_2 absorbs infrared Radiant Energy that would have reached the stratosphere, and the cycle repeats (Eisenhauer, 1998).

Scientist are well aware of the complexities of global warming but they are confident that the recent refreshment in measuring changes in the atmosphere, the new computer models being developed to predict long range changes in climate and the new efforts that these developments have inspired will soon combine to yield a detailed understanding of the global environment. It is this understanding that will in turn produce successful strategy for solving these problems at hand (Eisenhauer, 1998).

2.1.5 Atmospheric Chemistry and Feedback Of Green House Gases

All greenhouse gases except CO_2 and H_2O are removed from the atmosphere primarily by chemical processes within the atmosphere. Green house gases containing one or more H atoms (e.g. CH_4, HFCs and HCFCs), as well as other pollutants are removed primarily by the reaction with hydroxyl radicals (OH). The removal takes place in the troposphere, the lower part of the atmosphere, ranging from the surface up to 7 to 16km depending on latitude and season and containing 80% of the mass of the atmosphere. The green house gases N_2O, PFC, SF_6, CFCs and Halons do not react with OH in the stratosphere. These gases are destroyed in the stratosphere or above, mainly by solar ultraviolet radiation (UV) at short wavelength (<240nm), and are long-lived. Because of the time required to transport these gases to the region of the chemical loss, they have a minimum life time of 20 years. CO_2 is practically inert in the atmosphere and does not directly influence the chemical reaction, but it has a small insitu source from the oxidation of CH_4, CO and VOC (Prather, 1995 and Schimel,1996).

Troposphere OH^- abundance depend on abundances of NO, CH_4, CO, VOC, O_3 and H_2O plus the amount of solar UV (>30nm) that reaches the troposphere. As a consequence OH^- varies widely with geographical location, time of the day, and season. Likewise the local loss rates of all those gases reacting with OH^- also vary. Because of its dependence on CH_4 and other pollutants, tropospheric OH is expected to have changed since the preindustrial era and to change again for future emission scenarios. For some of these gases, other removal processes such as photolysis or surface uptake, are also important; and the total sink of the gas is obtained by integrating over all such processes. The chemistry of tropospheric O_3 is closely tied to that of OH and its abundance also varies with changing precursor emissions. The chemistry of the troposphere is also directly influenced by the stratospheric burden of O_3, climatic change in temperature (T) and humidity (H_2O) as well as by interactions between tropospheric aerosols and trace gases. Such compounds provide a "feedback" between the climate change induced by increasing green house gases and the concentration of these gases. Another feedbacks internal to the chemistry, is the impact of CH_4 on OH^- and hence its own loss. These feedbacks are expected to be important for tropospheric O_3 and OH^-. Such chemistry-chemistry or climate-chemistry coupling has been listed under "indirect effect" in the SAR (Prather, 1995 and Schimel, 1996)..

Stratospheric circulation and distribution of O_3 control the transport of the long-lived green house gases to regions of photochemical loss as well as the penetration of solar UV into the atmosphere. At the same time, many of these gases (e.g. NO2 and CFCs) supply ozone-depleting radicals (e.g. nitric oxide (NO and CL) to the stratosphere, providing a feedback between the gas and its loss rate. Another consequence of the observed stratospheric ozone depletion is that tropospheric photochemical activities are expected to have increased, altering troposphere OH^- and O_3. Climate change in the 21st century, including the radioactive cooling of the stratosphere by the increase levels of CO_2, is expected to alter stratospheric circulation and O_3 and the global mean loss rates of the long-lived gases. Some of these effects are discussed extensively in literature and are briefly considered here (Prather, 1995 and Schimel,1996).

The biosphere's response to global climate change has impact to the atmospheric composition of the 21st century. The anticipated change in climate (e.g. temperature, precipitation) will alter ecosystems and the "natural" background emission of trace gases. There is accumulating evidence that increased deposition of green house gases and elevated sulfur abundances have opposite influence on plant CO_2 uptake, NO_2 and O_3 (> 40ppb).The effects of deposition are detrimental to the increase in availability of atmospheric deposition and direct fertilization accelerate the emission of in-containing trace gases (NO,N_2O NH_3 and CH_4) as well as altering species diversity and biospheres' functioning. These complex interactions represent a chemistry-biosphere feedback that may alter greenhouse forcing (Prather, 1995 and Schimel,1996).

2.1.6 Atmospheric Lifetimes and Time-Scales

The global atmospheric lifetime (yr) characterizes the time required to turn over the global atmospheric burden. It is defined as the burden (Tg) divided by the mean global sink (Tg/yr) for a gas in steady state (i.e. with unchanging burden). The quantity was defined as both "lifetime" and "turnover time" in the SAR Bolin, R. (1973) Life time calculated in this manner is that when in steady state (i.e. source strength = sink strength) and the atmospheric burden of a gas equals the product of its lifetime and it emission. A further corollary of its steady – state lifetime for that emission pattern and the amount emitted (Prather, 1996).

The atmospheric lifetime is basically a scale factor relating.

(i) Constant emission (T g/yr) to a steady-state burden (T g).

(ii) An emission pulse (T g) to the time-integrated burden of that pulse (T g/yr). The lifetime is often additionally assumed to be a constant, independent of the sources; and it is also taken to represent the decay time (e-fold) of a perturbation.

These latter assumptions apply rigorously only for a gas whose local chemical lifetime is constant in space and time, such as for noble gas radon whose lifetime is fixed by the rate of its radioactive decay. In such a case the mean atmospheric lifetime equals the local lifetime, the lifetime that relates source strength to global burden is exactly the decay time of a perturbation (Prather, 1996).

This general application of the atmospheric lifetime breaks down for those greenhouse gases and pollutants whose chemical losses vary in space and time. NOx, for instance, has a local lifetime of < 1 day in the lower troposphere, but >5 days in the upper troposphere, and both times are less than the time required for vertical mixing of the troposphere. In this case, emission of NO_X into the upper troposphere will produce a larger atmospheric burden than the same emission into the lower troposphere. Consequently, the definition of the atmospheric lifetime of NO_X is not unique and depends on the location (and season) of its emissions. The same is true for any gas whose local lifetime is variable and on average shorter than about 0.5 year i.e. the decay time of a north-south difference between hemispheres representing one the longer time-scale for tropospheric mixing. The majority of green-house gases considered here have atmospheric lifetime greater than 2 years, much longer than troposphere mixing time; and hence their lifetime are not significantly altered in the location of the sources within the troposphere. When the lifetimes are reported for gases, it is assumed that the gases are uniformly mixed throughout the troposphere. This assumption is unlikely for gases with lifetime<1year, and reported values must be viewed by only approximation (Prather, *1996*).

2.1.7 Heat Transfer in the Atmosphere

Radiation energy from the sun passes through the Earth's atmosphere and reaches the Earth's surface. However, about 30% of the sun's total radiant energy reaching the Earth is reflected back into the space by the atmosphere, the clouds and the Earth surface. About 20 % is absorbed by the atmosphere, and the remaining 50 % is absorbed by the Earth's surface. Radiant energy absorbed by the Earth is changed into heat or infrared rays (invisible form of radiant energy that has a heat effect). Some of this heat absorbed by the Earth warms the atmosphere above it. This heat is transferred to the atmosphere by conduction (transfer of heat by direct contact between materials), convection (transfer of heat by movement of heated fluids-gases and liquid) and radiation (transfer of heat by emission of heat waves). Carbon dioxide and water vapour are gases in the atmosphere that help keep heat from being lost to space. They absorb heat from Earth and emit previously absorbed radiation) towards the earth. Like the plastic covering that prevents the escape of some of the infrared rays radiated from the soil, the earth's atmosphere keeps the earth warm (Pratther, *1996*).

2.1.8 Trace Gas Budget and Trends of Green House Gases

The "budget" of a trace gas consists of three quantities: its Global source, Global Sink and Atmospheric burden. The burden is defined as total mass of the gas integrated over atmosphere and related reservoirs, which usually include just the troposphere and stratosphere. The global burden (in Tg) and its trend (i.e. the differences between sources and sinks in Tg/yr) can be determined from atmospheric measurements and, for the long lived gases, are usually the best known quantities in the budgets. For short-lived, highly variable gases such as tropospheric O_3 and NOx, the atmospheric burden cannot be measured with great accuracy (Pratther, *1996*).

The global source strength is the sum of all sources, including emissions and insitu chemical production likewise, the sink strength (or global loss rate) can have several independent components (Pratther, *1996*).

The source strength (Tg/yr) for most green house gases is comprised of surface emission for synthetic gases where industrial production and emission are well documented, the source strength may be accurately known. For methane and N_2O, there are large, not well quantified, natural emissions (Pratther, *1996*).

Further, the anthropogenic emissions of these gases are primarily associated with agricultural sources that are difficult to quantify accurately. Considerable research has gone into identifying and quantifying the emissions from individual sources for CH_4 and N_2O. Such uncertainty in source strength also holds for synthetic gases with undocumented emissions. The source strength for tropospheric O_3 includes both a stratospheric influx and insitu production and is derived primarily from global chemical models (Pratther, *1996*).

The sink strength (Tg/yr) of long-lived greenhouse gases can be derived from a combination of atmospheric observations; laboratory experiments and models. The atmospheric chemistry models are based on physical principles and laboratory data, and include as constraints the observed chemistry of the atmosphere over the past two decades. For example, stratospheric loss rates are derived from models either by combining observed trace gas distributions with theoretically calculated loss frequencies or from the measured correction of the respective gas with a trace gas of known vertical flux. In such analysis there are a wide range of self-consistency checks. Mean global loss rates based on a prior modeling (e.g. the CH_4-lifetime studies from O_X comp described later) can be compared with empirically-based loss rates that are sealed from gas with similar loss processes that has well known emissions and atmospheric burden e.g. the AGAGE (Advanced Global Atmospheric Gases Experiment) calibration of mean troposphere OH using methyl chloroform (CH_3CCL_3). Our knowledge of the current budget of a greenhouse gas provides a key constraint in modeling its future abundance. (Prinn, 1999).

2.1.9 The Greenhouse Effects

In the mid-nineteenth century John Tyndall found that gases such as water vapour, carbon dioxide and methane strongly absorb infrared (thermal) radiation, while gases such as oxygen and nitrogen absorb very little (Houghton, 2002).

Theoretical and observed developments in physics during the twentieth century have shown that this is because different molecules absorb and re-emit infrared radiation of different wavelengths, depending on their molecular structure. For example, carbon dioxide strongly absorbs radiation with wavelengths of around 15 microns, and re-emits radiation at the same wavelengths (Houghton, 2002)

The atmosphere is made up of a number of different gases (Table 1,). The properties and abundance of these gases mean that, overall, the atmosphere is relatively transparent to visible light, which has short wavelengths, but tends to absorb infrared radiation, which has longer wavelengths (Figure 2.1, below) (Houghton, 2002)

Table 2.1 Average compositions of the dry atmosphere below 25km, Water vapour is additional to these figures, with its concentration varying significantly both spatially and cover time.

GAS	PERCENT VOLUME
Nitrogen	78.08%
Oxygen	20.95%
Argon	0.93%
Carbon Dioxide	~0.038%
Neon	0.00%
Helium	0.00%
Methane	~0.00018%
Hydrogen	0.00%
Nitrous Oxide	~0.00003%
Ozone	~0.000004%

(Houghton, 2002)

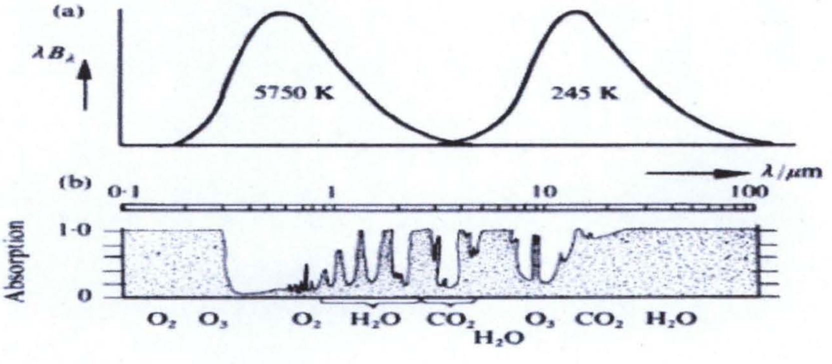

Figure 2.1 (a): Curves of black-body energy $B\lambda$ at wavelength λ at 5750K and 245K. The curves have been drawn of equal areas since it was integrated over the Earth's surface and all angles of the solar and terrestrial fluxes are equal.

Figure 2.1 (b) Absorption by atmospheric gases for a clear vertical column of atmosphere. Sources Houghton, (1998)

- Most of the energy radiated from the sun is concentrated in the visible (light) and near-visible parts of the electro-magnetic spectrum. These have short wavelengths and are able to pass through the atmosphere with little absorption. This is illustrated by the left hand side of Plate I (a,b). Overall, about 20% of solar energy that enters the Earth's atmosphere is reflected back into space by clouds and atmospheric particles and gases, and a further 10% is reflected by the Earth's surface (Figure 2.1a and 2.1b,). (Houghton, 2002)

- The rest of the sun's energy (about 70%) is absorbed by clouds and gases in the atmosphere and by land, water and vegetation at the surface. This energy warms the surface and is re-radiated in the form of heat (infrared radiation). As certain greenhouse gases, including water vapour, absorb infrared radiation, a significant proportion of this re-radiated energy is absorbed by the atmosphere before it escapes into space (Houghton, 2002)

- After gas molecules absorb radiation, they re-emit it in all directions. Some of the infrared radiation absorbed by gases in the atmosphere is therefore re-radiated out towards space and eventually leaves the atmosphere, but some is re-radiated back towards the Earth, warming the surface and at lower atmosphere (illustrated by the 'back radiation' term in Figure 2.2). This warming is known as the greenhouse effect and the gases that are responsible for it are known as greenhouse gases. (Houghton, 2002)

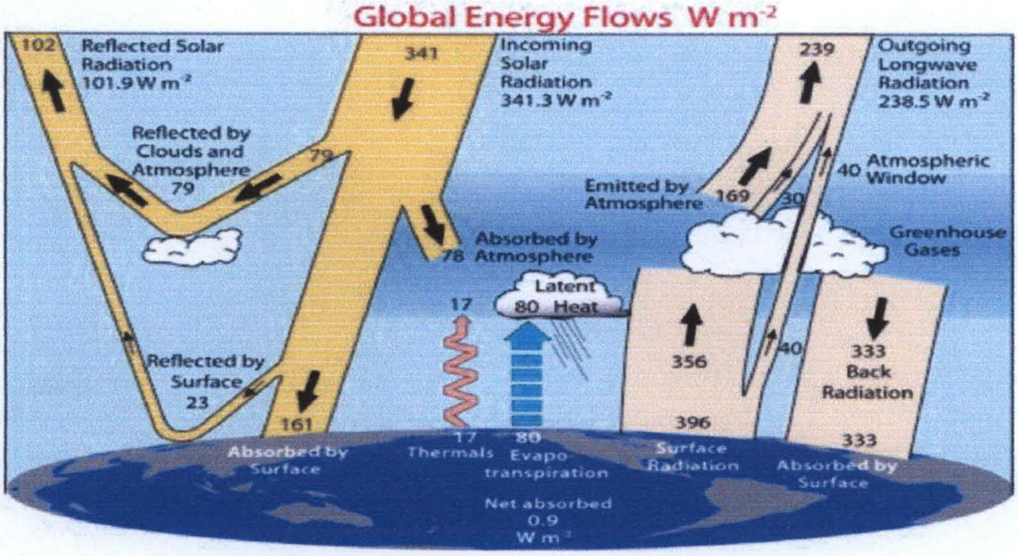

Figure 2.2: Global Average Energy Balance

2.1.10 Estimate of the Earth's Annual and Global Average Energy Balance - All Units Are (W/M^{-2})

- Without the greenhouse effect, the average temperature at the Earth's surface would be about 33°C colder (-18°C rather than 15°C). Other planets also have greenhouse effects: Mars has a relatively weak effect (its average surface temperature is about 3°C warmer than it would be without it), while Venus has a dense atmosphere rich in carbon dioxide, resulting in a strong greenhouse effect that keeps it relatively hot (its average surface temperature is about 470°C warmer than it would be without it) (Houghton, 2002).

- On Earth, water vapour is responsible for up to about 70% of the natural greenhouse effect. Carbon dioxide (CO_2), even though its concentration in the atmosphere is quite low, it is responsible for a significant proportion of the remaining greenhouse effect. This is because it is an efficient absorber of radiation with wavelengths that are emitted strongly by the Earth (i.e.

wavelengths near the peak of the Earth's 'emission spectrum', as shown by the right hand side in Figure (2.1a or b) and because CO_2 acts within the regions of the spectrum where water vapour effects are small (Houghton, 2002).

It is relatively straightforward to calculate the temperature change that will arise from a given change in the concentration of greenhouse gases in the climate system (assuming no feedbacks) because this 'radiative' warming is determined by basic physics. A doubling of the concentration of CO_2 in the atmosphere would cause a temperature rise of approximately 1°C. However this direct effect is modified by feedback processes in the atmosphere, as explained (Houghton, 2002).

2.1.11 Greenhouse Effect Model

The energy of light is transported in the air as electromagnetic radiation. Only a small fraction of this energy spectrum (visible light) can be seen by our eyes. Other parts are invisible and damage our skin (as ultraviolet light) or we can feel it as warm radiation (infrared radiation). The temperature in the atmosphere is considerably influence by the capability of the air molecules to absorb this radiation. Some of these models are discussed below (Sokolov, 2009).

2.2 Deterministic Models

Mathematical models in which outcomes are precisely determined through known relationships among states and events, without any room for random variations: In such models, a given input will always produce the same output, such as a known chemical reaction. In comparison stochastic models use range of values for variables in the form of probability distribution (Monte, 2011).

This strict deterministic prediction is based on the application of physical laws or first principle e.g. conservation of mass or conservation of energy with the outcome being completely determined by these laws. These predictions can be made by the absence of any similar observations. Deterministic prediction still contains errors and uncertainties. They can arise from uncertain initial condition, unforeseen forces and/or a poor application of the physical law (Sokolov, 2009).

2.2.1 Review of Mathematical Models of the Climate

To enable better understanding of the complex climate system, computer programs have been developed to model interactions of climate components. These general circulation models (GCMs) have been used extensively to understand climatic shifts observed in the past and to try to identify possible future responses of the climate system to changing conditions. Can the shifts occur over a short time, such as a decade or century? Will a shift be heralded by phenomena such as an increase in the frequency of El Niños and their surge of warm, western Pacific Ocean water toward South America? What are the different mechanisms of poleward heat transport that might provide the backbone of other climate states? These questions, and many others, indicate the complexity of current climate studies. Simple cause-and-effect arguments are usually not effective explanations in this arena. Complex computer models are practically the only tools available, so they are typically used to justify statements about climate and global dynamics (Sokolov, 2009).

For 20 years, climate-modeling researchers have been using some version of the Community Climate Model (CCM1) of the National Center for Atmospheric Research (NCAR). CCM1, which was produced in 1987, was operated on large serial supercomputers. Now, many of these researchers are using CCM2—a step forward that confidence to the reliability of these models as forecasting tools for the future climate resulting from increasing greenhouse gases (Sokolov, 2009).

CCM^2, the most recent code in a series of climate models developed by NCAR, captures the intricate interactions of the physical processes outlined here. This climate model, which is available to academic and industrial research users, simulates the time-dependent response of the climate system to the daily and seasonal variation of the solar input and of sea surface temperatures. For the past 10 years and into the foreseeable future, these models form the basis of a broad range of climate research and scenario testing used in support of decision makers who formulate national energy and environmental policies (Sokolov, 2009).

2.2.2 An Elementary Climate Model

A crude model of the temperature of the earth is based on the conservation of energy, and can readily be built. One simply assumes the system to be a balance requiring that incoming radiation equals outgoing radiation. The following assumptions are made that the earth emits radiation as a perfect black body, there is no atmosphere or ocean and the earth reflects in coming solar radiation with a uniform albedo. Such a crude model would lead to a prediction of the earth being- 18°C, which is rather cold (Sokolov, 2009).

2.2.3 An Elementary Climate Model a Simple Atmosphere

One can easily extend such a model by including an ideal, single layer model of the atmosphere that covers the planet. In addition to the initial assumptions made, one needs to further assume that the atmosphere is single perfectly well mixed layer, the atmosphere is a perfect grey body with a uniform emissivity and absorptivity, and

and land surface components. GCMs and global climate models are widely applied for weather forecasting, understanding the climate, and projecting climate change. Versions were designed for decades to century and climate applications were originally created by Syukuro Manabe and Kirk Bryan in 1973 at the Geophysical Fluid Dynamics Laboratory in Princeton. New Jersey. These computationally intensive numerical models are based on the integration a variety of fluid dynamics, chemical, and sometimes biological equations (Sokolov, 2009).

When developed a mathematical model which could realistically depict monthly and seasonal patterns in the troposphere, which became the first successful climate model, following Phillip's work, several groups began working to create general circulation models. The first general circulation climate model that combined both oceanic and atmospheric process was developed in the late 1960s at the NDAA Geophysical fluid Dynamics Laboratory, and by the early 1980s, the United States' National Centre for atmospheric Research had developed the community Atmosphere model; this model has been continuously refined into the 2000s. In 1986, efforts began to initialize and model soil and vegetation types, which led to more realistic forecasts. Coupled with ocean atmosphere climate models such as the Hadley Centre for climate prediction and Research's Had CM_3 model are currently being used as inputs for climate change studies. The importance of gravity waves was neglected within these models until the mid 1980. Gravity and waves are required within global climate models in order to properly simulate region scale circulation, though their broad spectrum makes their incorporation complicated (Pope, 2008).

2.2.4.1Emission Scenarios

For the six SRES maker scenarios, gave a "best estimate" of global mean temperature increase (2090 - 2099) relative to the period 1980 – 1999 that ranged from 1.8°C to 4.0°C. Over the same time period, the "likely" range (greater than 66% probability, based on expert judgments) for these scenarios was for a global mean temperature increase of between 1.1 and 6.4°C. Pope, (2008) described a study where climate change projections were made using several different emission scenarios. In a scenario where global emissions start to decrease by 2010 and then decline at a sustained rate of 3 % per year, the likely global average temperature increase was predicted to be 1.7 ° C above pre – industrial levels by 2050, rising to around 2 ° C by 2100. In a projection designed to simulate a future where no effort are made to reduce global emissions, the likely rise in global average temperature was predicted to be 5.5 ° C by 2100. A rise as high as 7 ° C was thought possible but less likely (Sokolov, 2009).

Sokolov, (2009) examined a scenario designed to simulate a future where there is no policy to reduce emissions. In there integrated model, this scenario resulted in a median warming over land (2090 – 2099 relative to the period 1980 - 1999) of 5.1 ° C. Under the same emission scenario but with different modeling of future climate, the median warming was 4.1 ° C.

2.2.4.2 Accuracy of Models That Predict Global Warming

AOGCM represents the pinnacle of complexity in climate models and internalize as many processes as possible. However, they are still under development and uncertainties remain. They may be coupled to models of other processes, such as the carbon cycle, so as to better model feedback effects. Most recent simulations show "plausible" agreement with the measured temperature anomalies over the past 150 years, when forced by observed changes in green house gases and aerosols, but better agreement is achieved when natural forcings are also included (Sokolov, 2009).

No model – whether a wind tunnel model for designing air craft, or or a climate model for projecting global warming – perfectly reproduces the system being modeled. Such inherently imperfect models may nevertheless produce useful results. In this context, GCMs are capable of reproducing the general features of the observed global temperature (Sokolov, 2009).

A debate over how to reconcile climate model predictions that upper air (tropospheric) warming should be greater than surface warming, with observations some of which appeared to show otherwise now appears to have been resolved in favor of the models ,following revisions to the data (Sokolov, 2009).

The effects of clouds are a significant area of uncertainty in climate models. Clouds have competing effects on the climate. One of the roles that cloud play in climate is in cooling the surface by reflecting sunlight back to the space; another is warming by increasing the amount of infrared radiation emitted from the atmosphere to the surface. In the 2001 IPCC report on climate change, the possible changes in cloud cover were highlighted as one of the dominant uncertainties in predicting future climate change (Sokolov, 2009),

2.2.5 Computations Involved

Climate models use quantitative methods to simulate the interactions of the atmosphere, oceans, land surface, and ice. They are used for a variety of purposes from study of the dynamics of the climate system to projections of future climate (Sokolov, 2009).

All climate models have taken account of incoming energy as short-wave (near) infrared, as well as outgoing energy as long wave (Far) infrared electromagnetic radiation from the earth. Any imbalance results in a change in temperature. (Sokolov, 2009).

The most talked-about models of recent years have been those relating temperature to emissions of carbon dioxide (see greenhouse gas). These models project an upward trend in the surface temperature record, as well as a more rapid increase in temperature at higher attitudes. Three (or more property, four since time is also considered), dimensional GCM's discretise the equations for fluid motion and energy transfer and integrate these over time. They also contain parameterizations' for process – such as convection – that occur on scales too small to be resolved directly (Sokolov, 2009).

Atmospheric GCMs (AGCMs) model the atmosphere and impose seas surface temperature as boundary conditions. Coupled atmosphere – ocean GCMs (AOGCMs) e.g. HadCM3, EdGCM, GFDL CM2.X, ARPEGE-Climate combine the two models. Models can range from relatively simple to quite complex.

- A simple radiant heat transfer model that treats the earth as a single point and averages outgoing energy. Examples include (Sokolov, 2009);
- This can be expanded vertically (radiative-convective models), or horizontally (Sokolov, 2009).
- Finally, (coupled) atmosphere-ocean-sea, ice global climate models discrete and solve the full equations for mass and energy transfer and radiant exchange (Sokolov, 2009).

2.3 Physical Impacts

This section describes some physical impacts of climate of change. For some of these physical impacts, their effects on social and economic systems are also noted by (Smith,and Schneider, 2009).

2.3.1 Effects of Weather

Increasing temperature is likely to lead to increasing precipitation by the effects on storms are less clear. Extra tropical storms partly depend on the temperature gradient, which is predicted to weaken in the northern hemisphere as the polar region warns more than the rest of the hemisphere by (Smith and Schneider 2009).

2.3.2 Extreme Weather

The predicted that in the future, over most land areas, the frequency of warm spells or heat waves would very likely increase. Other changes are listed below.

- Increase areas will be affected by draught
- There will be increased intense of extreme tropical cyclone activity
- There will be increased incidences of extreme high seas level (excluding tsunamis).

2.3.3 Local Climate Change

Regional effects of global warming vary in nature. Some are the result of generalized global changes, such as rising temperature, resulting in local effects, such as melting ice in other cases; a change may be related to a change in a particular ocean current or weather system. In such cases, the regional effect may be disproportional and will not necessarily follow the global trend (Smith, and Schneider 2009).

There are three ways in which global warming will make changes to original climate. Melting or forming ice, changing the hydrological cycles (of evaporation and precipitation and changing current in the oceans and air flows in the atmosphere. The coast can also be considered a region, and will suffer severe impacts from sea level rise. (Sokolov, 2009)

2.3.4 Biogeochemical Cycles

Climate change may have an effect on carbon cycle in an interactive "feedback" process. A feedback exists where an initial process triggers changes in a second process that in turn influences the initial process. A positive feedback intensifies the original process, and a negative feedback reduces it. Models suggest that the interaction of the climate system and carbon cycle is one where the feedback effect is positive (Schneider, 2007).

Using the A_2SRES emission scenario Schneider, (2007) found that this effect led to additional warming by 2010, relative to the 1990-2000 period, of 0.1 to 1.5 $^{\circ}$ C. This estimate was made in the IPCC Forth Assessment Report of 1.1 to 6.4 $^{\circ}$ C account for this feedback effect. On the other hand with medium confidence, Schneider, (2007) commented that additional releases of GHGS were possible from permafrost, peat lands, wetlands, and large stores of marine hydrates at high latitudes (Schneider, 2007).

Amplitude: Amplitude is a parameter characterizing the annual variation of soil temperature around an average value. If the variation in temperature within a day is averaged out over many years, the annual amplitude is one-half the difference between this annual averaged maximum and annual averaged minimum temperatures within a year.

Damping depth: Damping depth is a constant characterizing the decrease in amplitude with an increase in distance from the soil surface. It is defined as $(2D_h/\omega)^{1/2}$, where D_h is the thermal diffusivity and ω is the frequency of a temperature fluctuation. For annual fluctuation $\omega = 2\pi/365\ d^{-1}$.

Thermal diffusivity: Thermal diffusivity is the change in temperature produced in a unit volume by the quantity of heat flowing through the volume in unit time under a unit temperature gradient. It can be calculated from thermal conductivity and volumetric heat capacity.

Time lag: Time lag is the number of days from an arbitrary starting date to the occurrence of the minimum temperature in a year.

2.5 Physical Properties Of Soil

The physical properties of a soil are those characteristics which can be seen with the eye or felt between the thumb and fingers. They are the result of soil parent materials being acted upon by climatic factors (such as rainfall and temperature), and affected by topography (slope and direction, or aspect) and life forms (kind and amount, such as forest, grass, or soil animals) over a period of time. A change in any one of these influences usually results in a difference in the type of soil formed. Important physical properties of a soil are color, texture, structure, drainage, depth, and surface features (stoniness, slope, and erosion).

The physical properties and chemical composition largely determine the suitability of a soil for its planned use and the management requirements to keep it most productive. To a limited extent, the fertility of a soil determines its possible uses, and to a larger extent, its yields. However, fertility level alone is not indicative of its productive capacity, since soil physical properties usually control the suitability of the soil as growth medium. Fertility is more easily changed than soil physical properties (Bristow *et al, 2001).*

2.5.1Color

When soil is examined, color is one of the first things noticed. It indicates extremely important soil conditions. In general, color is determined by: (1) organic matter content, (2) drainage conditions, and (3) degree of oxidation (extent of weathering). Surface soil colors vary from almost white, through shades of brown and gray, to black. Light colors indicate a low organic matter content and dark colors can indicate a high content. Light or pale colors in the surface soil are frequently associated with relatively coarse texture, highly leached conditions, and high annual temperatures. Dark colors may result from high water table conditions (poor drainage), low annual temperatures, or other conditions that induce high organic matter content and, at the same time, slow the oxidation of organic materials. However, soil coloration may be due to the colors imparted by the parent material. Shades of red or yellow, particularly where associated with relatively fine textures, usually indicate that subsoil material has been incorporated in the surface layer.

Subsoil colors, in general, are indications of air, water, and soil relationships and the degree of oxidation of certain minerals in the soil. Red and brown subsoil colors indicate relatively free movement of air and water allowed by the soil. If these or other bright colors persist throughout the subsoil, aeration is favorable. Some subsoils that are mottled (have mixed colors), especially in shades of red and brown, are also well-aerated.

Yellow-colored subsoils usually indicate some drainage impediment. Most mottled subsoils, especially those where gray predominates, have too much water and too little air (oxygen) much of the time. The red-to-brown color of subsoils comes from iron coatings under well-aerated conditions. In wet soils with low oxygen levels, the iron coatings are chemically and biologically removed, and the gray color of background soil minerals shows.

2.5.2 Texture

Texture refers to the relative amounts of differently sized soil particles, or the fineness/coarseness of the mineral particles in the soil. Soil texture depends on the relative amounts of sand, silt, and clay. In each texture class, there is a range in the amount of sand, silt, and clay that class contains.

The coarser mineral particles of the soil are called sand. These particles vary in size. Most sand particles can be seen without a magnifying glass. Sand particles feel gritty when rubbed between the thumb and fingers. Relatively fine soil particles that feel smooth and floury are called silt. When wet, silt feels smooth but is not slick or sticky. When dry, it is smooth, and if pressed between the thumb and finger, will retain the imprint. Silt particles are so fine that they cannot usually be seen by the unaided eye and are best seen with a microscope. Clays are the finest soil particles. Clay particles can be seen only with the aid of a very powerful (electron) microscope. They feel extremely smooth when dry, and become slick and sticky when wet. Clay will hold the form into which it is molded. Soils high in clay content often show pronounced surface cracking when dried.

Soil textural classes take their names from the particle size categories (sand, silt, and clay) and also from the category called loam. Loam is a textural class of soil that has moderate amounts of sand, silt, and clay. Loam contains approximately 7% to 27% clay, 28% to 50 % silt, and 23% to 53% sand. It is smooth to the touch when dry, but when moist, it becomes somewhat slick/sticky.

Most surface soils fall into five general textural classes. Each class name indicates the size of the mineral particles that are dominant in the soil. Intermediate texture soils are called loams. Texture is determined in the field by rubbing moist-to-wet soil between the thumb and fingers. These observations can be checked in the laboratory by mechanical analysis which separates particles into clay, silt, and various-sized sand groups.

Estimating Soil Texture Soils can be classified by their texture or particle size distribution. Practically speaking, a soil particle must pass through a 2mm sieve (No. 10, US) to be called a soil particle (Bristow *et al*, 2001).

Soil particles are either organic or inorganic. Texture is used only to describe the distribution of the inorganic fraction. In the laboratory, the organic matter is either removed prior to mechanical analysis. In the field, organic matter is disregarded when the texture is determined.

An estimation of texture can be made in the field by using the following method:

1. Place about a tablespoon of soil in the palm of your hand.
2. Mix it with water and form a moist ball. The soil is at the correct consistency when the ball does not leave soil on the palm of your hands when it is rolled around (the consistency of modeling clay.)
3. Press the moistened soil ball between the thumb and forefinger in an attempt to form a ribbon with the soil. As the thumb and forefinger are pressed together the soil will extrude forming the ribbon. The motion should be repeated several times to test the cohesiveness of the ribbon, attempting to form a continuous ribbon.
4. Ribbons can be classed into three broad categories:
    - Good Ribbon —The ribbon does not break and has very little cracking along the sides.
    - Medium Ribbon —The sides of the ribbon crack deeply and eventually the ribbon will break and fall off.
    - Poor Ribbon —No ribbon formed (no cohesiveness) or the ribbon breaks with the first applied pressure and does not cohere.
5. The sample is then further wetted and mashed between the thumb and forefinger. This is to determine the amount of grittyness or smoothness. The soil should be wet enough to feel individual particles. If the soil feels like sand, then it is called "gritty." If it feels like flour then it is called "smooth." There is a category between the two that is both smooth and gritty but no descriptive term is used (Bristow *et al,* 2001).

2.5.3 Soil thermal properties

The **thermal properties of soil** are a component of soil physics that has found important uses in engineering, climatology and agriculture. These properties influence how energy is partitioned in the soil profile. While related to soil temperature, it is more accurately associated with the transfer of heat throughout the soil, by radiation, conduction and convection.

The main soil thermal properties are:

- Volumetric heat capacity, SI Units: $J.m^{-3}\cdot K^{-1}$
- Thermal conductivity, SI Units: $W.m^{-1}\cdot K^{-1}$
- Thermal diffusivity, SI Units: $m^{2}\cdot s^{-1}$

2.5.4 Measurement

It is hard to say something general about the soil thermal properties at a certain location because these are in a constant state of flux from diurnal and seasonal variations. Apart from the basic soil composition, which is constant at one location, soil thermal properties are strongly influenced by the soil volumetric water content, volume fraction of solids and volume fraction of air. Air is a poor thermal conductor and reduces the effectiveness of the solid and liquid phases to conduct heat. While the solid phase has the highest conductivity it is the variability of soil moisture that largely determines thermal conductivity. As such soil moisture properties and soil thermal properties are very closely linked and are often measured and reported together. Temperature variations are most extreme at the surface of the soil and these variations are transferred to sub surface layers but at reduced rates as depth increases. Additionally there is a time delay as to when maximum and minimum temperatures are achieved at increasing soil depth (sometimes referred to as thermal lag (Bristow *et al, 2001).*

One possible way of assessing soil thermal properties is the analysis of soil temperature variations versus depth Fourier's law,

$$Q = -\lambda dT/dz$$

where Q is heat flux or rate of heat transfer per unit area $J \cdot m^{-2} \cdot s^{-1}$ or $W \cdot m^{-2}$, λ is thermal conductivity $W \cdot m^{-1} \cdot K^{-1}$; dT/dz is the gradient of temperature (change in temp/change in depth) $K \cdot m^{-1}$.

The most commonly applied method for measurement of soil thermal properties is to perform in-situ measurements, using Non-Steady-State Probe systems, or Heat Probes (Bristow *et al, 2001).*

2.5.5 Remote sensing

Remote sensing from satellites, aircraft has greatly enhanced how the variation in soil thermal properties can be identified and utilized to benefit many aspects of human endeavor. While remote sensing of reflected light from surfaces does indicate thermal response of the topmost layers of soil (a few molecular layers thick), it is thermal infrared wavelength that provides energy variations extending to varying shallow depths below the ground surface which is of most interest. A thermal sensor can detect variations to heat transfers into and out of near surface layers because of external heating by the thermal processes of conduction, convection, and radiation. Microwave remote sensing from satellites has also proven useful as it has an advantage over TIR of not being effected by cloud cover (Bristow *et al, 2001)*.

The various methods of measuring soil thermal properties have been utilized to assist in diverse fields such as; the expansion and contraction of construction materials especially in freezing soils, longevity and efficiency of gas pipes or electrical cables buried in the ground, energy conservation schemes, in agriculture for timing of planting to ensure optimum seedling emergence and crop growth, measuring greenhouse gas emissions as heat effects the liberation of carbon dioxide from soil. Soil thermal properties are also becoming important in areas of environmental science such as determining water movement in radioactive waste and in locating buried land mines (Bristow *et al, 2001)*.

CHAPTER THREE

3.0 MATERIALS AND METHODS

3.1 Model Development

3.1.1 Assumptions for The Model Development.

In order to simplify the model, impose limitations, make the process of obtaining solutions and the predictive nature of the model more accurate, the following assumptions were made;

(1) Energy is assumed to come from the sun and radiate from the earth surface.

(2) The albedo (Reflection Coefficient) and emissivity of the Earth surface is assumed to be the same in the area under the study.

(3) Energy transported in the atmosphere is controlled by convection.

(4) It is assumed that 0.009 of energy is convectively transferred from the sun to the atmosphere and 0.00000006 of energy is radiated from the earth surface.

(5) The minimum temperature for materials to radiate heat is assumed be 15 ^{0} C.

Subsoil colors, in general, are indications of air, water, and soil relationships and the degree of oxidation of certain minerals in the soil. Red and brown subsoil colors indicate relatively free movement of air and water allowed by the soil. If these or other bright colors persist throughout the subsoil, aeration is favorable. Some subsoils that are mottled (have mixed colors), especially in shades of red and brown, are also well-aerated.

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2.3.4 Biogeochemical Cycles

Climate change may have an effect on carbon cycle in an interactive "feedback" process. A feedback exists where an initial process triggers changes in a second process that in turn influences the initial process. A positive feedback intensifies the original process, and a negative feedback reduces it. Models suggest that the interaction of the climate system and carbon cycle is one where the feedback effect is positive (Schneider, 2007).

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2.3.4.1 Ground-level ozone

With high confidence, Confalonieri, (2007) projected that climate change would increase cardio respiratory morbidity and mortality associated with ground-level ozone. Ground level ozone is both naturally occurring and is the primary constituent of urban 5m0g (Confalonier, 2007).

Ozone in 5mOg is formed through chemical reactions involving nitrogen oxides and other compounds. The reaction is a photochemical reaction, meaning that it involves electromagnetic radiation, and occurs in the presence of bright sun shine and high temperatures. Exposure to elevated concentrations of ozone is associated hospital admission for pneumonia, chronic obstructive pulmonary disease, asthma, allergic rhinitis and other diseases and with premature mortality (Confalonieri, 2007).

Background levels of ground-level ozone have risen since pre-industrial times because of increasing emissions of methane, carbon monoxide and nitrogen oxides (Confalonieri, 2007).

2.3.4.2 Cold waves

It is expected that climate change will bring some health benefits Confalonieri, (2007). It is expected that these benefits will be outweighed by negative climate change effects. Cold-waves continues to be a problem in northern latitudes, where very low temperatures can be reached in a few hours and extend over long periods Confalonieri, (2007). Reductions in cold-deaths due to climate change are projected to be greater than increases in heat related deaths in the UK. (confalonieri, 2007).

2.3.5 Water Resources

In a literature assessment, Kundzewicz, (2007) concluded, with high confidence, that;

- The negative impacts of climate change on fresh water systems outweigh the benefits. All of the regions assessed in the IPCC fourth Assessment Report (Africa, Asia, Australia and New Zealand, Europe, Latin America, North America, Polar regions (Artic and artatic), and small island) showed on overall net negative impact of climate change on water resources and fresh water ecosystems (Confalonieri, 2007).
- Semi-arid and arid areas are particularly exposed to the impacts of climate change on fresh water. With very high confidence, it was judged that many of these areas, e.g. the Mediterranean basin, western USA, Southern Africa, and North-eastern Brazil, would suffer a decrease in water resources due to climate change (Sokolov, 2009).

2.3.5 Migration and Conflict

An argument can be made that rising ethnic conflicts may be linked to competition over natural resources that are increasingly scarce as a result of climate change. According to a literature assessment by Wilbanks, (2007), other factors need to be taken into according. It was suggested that major environmentally influenced conflicts in Africa have more to do with the relative abundance of resources e.g. oil and diamonds, than with resources scarcity. On this basis, Wilbanks, (2007) suggested that predictions of future conflicts due to climate change should be viewed with caution (Wilbanks, 2007).

With high confidence, Schneider, (2007) predicted that stressed such as increased drought, water shortages, and riverine and coastal flooding would affect many local and regional populations. With medium confidence, it was predicted that these stresses would lead, in some eases, to relocation within or between countries. This might have the effect of exacerbating conflicts and possibly impose migration pressure. Zhang *et al, (2003)* also concluded that climate change drives conflict (Wilbanks, 2007).

2.4 Soil Temperature Variations with Time and Depth

Soil temperature fluctuates annually and daily affected mainly by variations in air temperature and solar radiation. The annual variation of daily average soil temperature at different depths can be estimated using a sinusoidal function (Hillel, 1982; Marshall and Holmes, 1988; Wu and Nofziger, 1999). This program estimates daily soil temperatures and displays these values as functions of time or depth for user defined input parameters (Hillel, 1982; Marshall and Holmes, 1988; Wu and Nofziger, 1999).

2.4.1 Model Description

The annual variation of daily average soil temperature at different depths is described with the following sinusoidal function (Hillel, 1982):

where $T(z,t)$ is the soil temperature at time t (d) and depth z (m), T_a is the average soil temperature (°C), A_0 is the annual amplitude of the surface soil temperature (°C), d is the damping depth (m) of annual fluctuation and t_0 is the time lag (days) from an arbitrary starting date (taken as January 1 in this software) to the occurrence of the minimum temperature in a year.

$$T(z,t) = T_a + A_0 e^{-z/d} \sin\left[\frac{2\pi(t-t_0)}{365} - \frac{z}{d} - \frac{\pi}{2}\right]$$

The damping depth is given by $d = (2D_h/\omega)^{1/2}$, where D_h is the thermal diffusivity and $\omega = 2\pi/365\ d^{-1}$.

2.4.2 Assumptions and Simplifications

The sinusoidal temperature model was derived by solving the following partial differential equation (Hillel, 1982; Marshall and Holmes, 1988):

where $T(z,t)$ is the soil temperature at time t and depth z and D_h is the thermal diffusivity.

$$\frac{\partial T(z,t)}{\partial t} = D_h \frac{\partial^2 T(z,t)}{\partial z^2} \quad T(0,t) = T_a + A_0 \sin\left[\frac{2\pi(t-t_0)}{365}\right]$$

The following assumptions are employed in the derivation of the temperature model:

1. A sinusoidal temperature variation at the soil surface z = 0. That is

where T_a is the average soil temperature, A_0 is the amplitude of the annual temperature function, t_0 a time lag from an arbitrary starting date (selected as January 1 in this software) to the occurrence of the minimum temperature in a year.

1 At infinite depth, the soil temperature is constant and is equal to the average soil temperature.

The thermal diffusivity is constant throughout the soil profile and throughout the year.

2.4.2 Frequently Asked Questions

1. *How well does the model predict soil temperatures?* Figure 1 compares measured and predicted soil temperatures at 4 depths for a site (located at 36.87 degrees North and 115.57 degrees East) in Hebei Province, China (Wu and Nofziger, 1999). The parameters in the model in this case were obtained from measured soil surface temperatures, clay content and average water content of each layer. Clearly the model does a good job predicting the mean daily temperatures. Figure 1. Measured mean and predicted soil temperatures at four depths based on measured soil surface temperatures.

2. *Can I use air temperatures to estimate soil temperatures?* Figure 2 compares the measured and predicted soil temperatures for the same site when model parameters are obtained from air temperatures instead of measured soil surface temperatures (Wu and Nofziger, 1999). The model consistently underestimates soil temperatures by about 2 degrees Celsius. These data and those of others suggest that good estimates of temperatures under bare soils can be obtained using by simply increasing the maximum and minimum air temperatures by 2 degrees when defining the model parameters. The

correction for soils that are not bare will likely be less since those soil temperatures are somewhat less due to shading from the plants. Figure 2: Comparison of measured and predicted soil temperatures at four depths based upon air temperatures at the site(Hillel, 1982; Marshall and Holmes, 1988; Wu and Nofziger, 1999).

3. *What is the effect of soil wetness on soil temperature distributions?* The thermal diffusivity of the soil is the ratio of the thermal conductivity of the soil to the volumetric heat capacity of the soil. The conductivity and volumetric heat capacity increase with water content so the diffusivity is also dependent upon soil water content. For mineral soils, the thermal diffusivity increases with water content at low water contents and then gradually decreases with increasing water contents at high water contents. This is illustrated in the Figure 3 for three clay contents and for three bulk density values. The volumetric heat capacity fot the three bulk densities is shown in Figure 2.3(Hillel, 1982; Marshall and Holmes, 1988; Wu and Nofziger, 1999).

3.2 The Diagram of the Shape To be Model

The experiment was performed using a shoe box as shown in Figure 3.1; it has a size of 5cm by 12cm and contained the soil and thermometer.

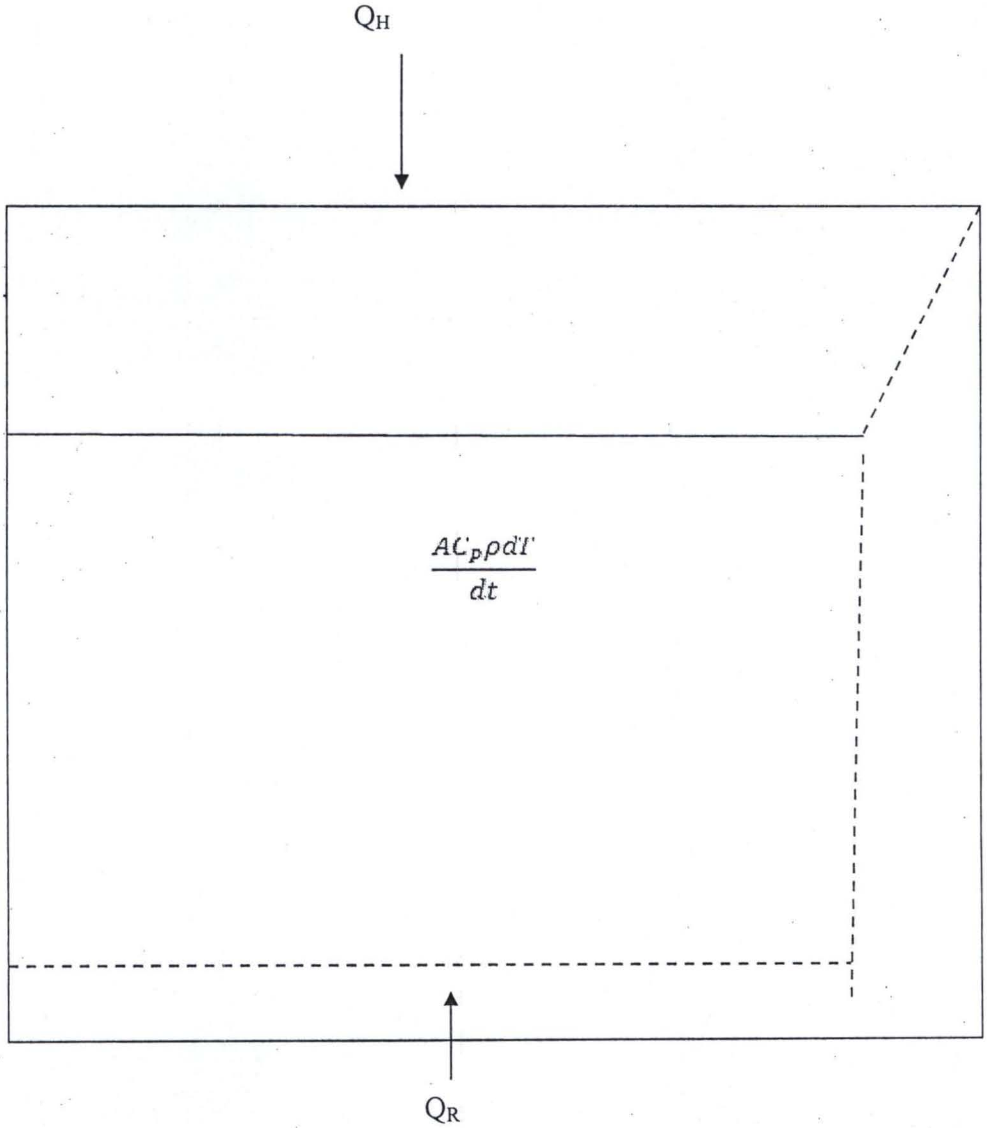


Figure 3. 1. Representation of the shoe box for performing the experiment

With respect to Figure 3.1, the following parameters are hereby defined

$T_m - Minimum\ temperature\ of\ the$ earth (15 ° C)

Q_H - Heat transfer into the atmosphere (joules)

Q_R – Heat radiation by the earth surface (joules)

C_P – Specific Heat capacity of the gases (j/kg ° C)

ρ – Density of the gases (kg/m^3)

dt – The change in time (minute)

dT – Change in temperature of the atmosphere (° C)

A – Area of the atmosphere under the study (m^2)

α –Albedo is the reflextivity of the earth

h – Heat transfer coefficient of the atmosphere (J/m^2 Min ° C)

ε – Emissivity and δ- Boltman Stefan's constant has already been defined.

(J/m^{2}Min ° C^4)

3.3 Model Development

$$\begin{array}{c}\text{Heat Transfer within}\\ \text{the Shoe box}\end{array} = \begin{array}{c}\text{Heat Transfer From}\\ \text{The Sun}\end{array} + \begin{array}{c}\text{Heat Radiation from}\\ \text{The Materials}\end{array} \qquad 3.1$$

Heat Transfer within the shoe box = $\frac{AC_P\rho dT}{dt}$ 3.2

Heat Transfer from the sun through convection= $Q_H = hAT_m$ 3.3

Heat Radiation by the material into the shoe box = $Q_R = 4A\varepsilon\delta T_m^4$ 3.4

Hence, 3.2, 3.3 and 3.4 into 3.5 gives

3.4 Modeled Equation

$$\frac{AC_p \rho dT}{dt} = Q_H + Q_R \qquad 3.5$$

Equation 3.5 can also be written as

$$A\rho C_{p\frac{dT}{dt}} = hAT_m + 4Aa\varepsilon\delta T_m^4 \qquad 3.6$$

That is

$$A\rho C_p dT = hAT_m dt + 4A\varepsilon\delta T_m^4 dt \qquad 3.7$$

$$dT = \frac{hT_m dt}{\rho C_p} + \frac{4\varepsilon\delta T_m^4}{\rho C_p} dt \qquad 3.8$$

Therefore, integrating

$$\int_{T_0}^{T} dT = \int_{t_0}^{t} (\frac{hT_m}{\rho C_p} + \frac{4\varepsilon\delta T_m^4}{\rho C_p}) dt \qquad 3.9$$

Finally

$$\Delta T = \left(\frac{hT_{m+4\varepsilon\delta T_m^4}}{\rho C_p} \right) \Delta t \qquad 3.10$$

Equation 3.10 will be used to compute the Simulated Temperature

3.5 Experimental Work

The main aim of the experiments conducted was to demonstrate the greenhouse effect. The materials, source and experimental procedures used are detailed in the following sections.

3.5.1 Materials and Equipment

Table 3.1: The materials and equipment used

S/N	MATERIAL	SOURCE
1	Two Shoe boxes	A.A Shoes Suleja
2	Rulers	Eze Book Shop
4	Two Thermometers	Laboratory
5	Colorless Plastic Food Wrap	Suleja Plastics
6	Timer	Laboratory
7	Organic Soil	Dantata & Sawoe Abuja
8	Laterite	Dantata & Sawoe Abuja
9	Sandy Soil	Gurrara River
10	Chippings	Dantata & Sawoe Abuja
11	Charcoal	Madam Ugwu Charcoal Seller

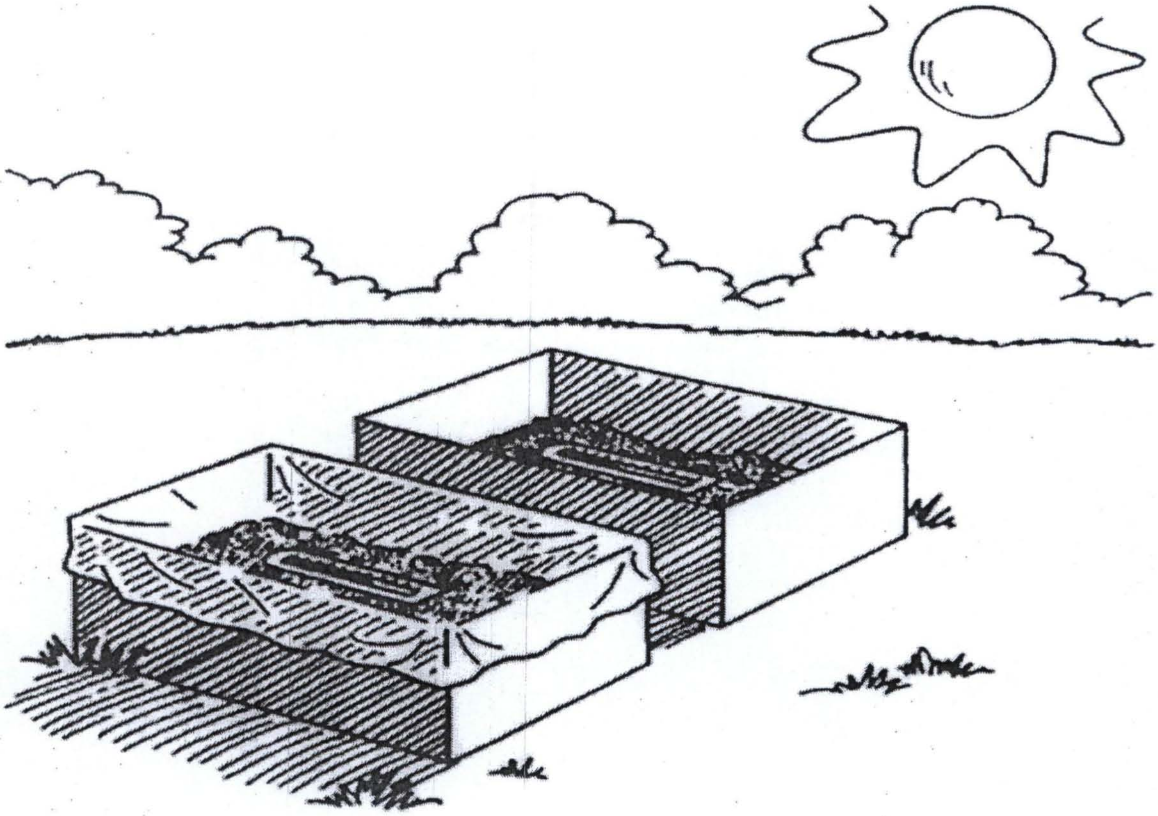


Figure 3.2: The Diagram of the Experiment to Demonstrate Green House Effect

3.5.2 Experimental Procedure

Two shoe boxes of 5cm by 12cm each were obtained. The bottom of each box was covered with 2 inches (5mm) the same material. The thermometers were placed on the surface of the material in each box. The top of one box was covered with a single layer of transparent plastic sheet and the other box was left uncovered and readings were taken from both thermometers in every 15 minutes for one hour. The boxes were placed side by side in a sunny place out doors. The thermometer readings were taken. And the temperatures of the thermometers in the covered and uncovered shoe boxes were recorded.

3.5.3 Experimental Temperature Change

The experimental temperature change is calculated as the difference between the temperature reading of the covered shoe boxes from the temperature reading of the uncovered one.

$$\Delta T_e = T_{(covered)} - T_{(uncovered)}$$

The above experiment was repeated using laterite, sharp sand, charcoal, organic soil and chippings and the results are presented in chapter four.

3.5.4 Modeling and Simulation

The modeled equation 3.10 was simulated by inputting the parameters of atmospheric gases in the Appendix A obtained from chemical engineering hand book and the result temperature change of green house effect from simulated result will be compared with that from experimental result.

CHAPTER FOUR

4.0 ` Results and Discussion of Results

The following results were obtained from the experiments and simulation. The first set of the Table represents the temperature change while the set of graph represents the temperature change with respect to time, the graph also displays the slope or rate of change.

Table 4.1: Experimental and simulated temperature change with organic soil

t/minutes	$T_{(covered)}$ °C	$T_{(uncovered)}$ °C	ΔT_e °C	ΔT_S °C	ΔT_{e-s} °C
15	43	40	3.000	2.279	0.721
30	48	42	6.000	4.588	1.412
45	53	44	9.000	6.837	2.163
60	56	45	11.000	9.117	1.883

ΔT = Temperature at the end of time (t) – Initial Temperature at time (t).

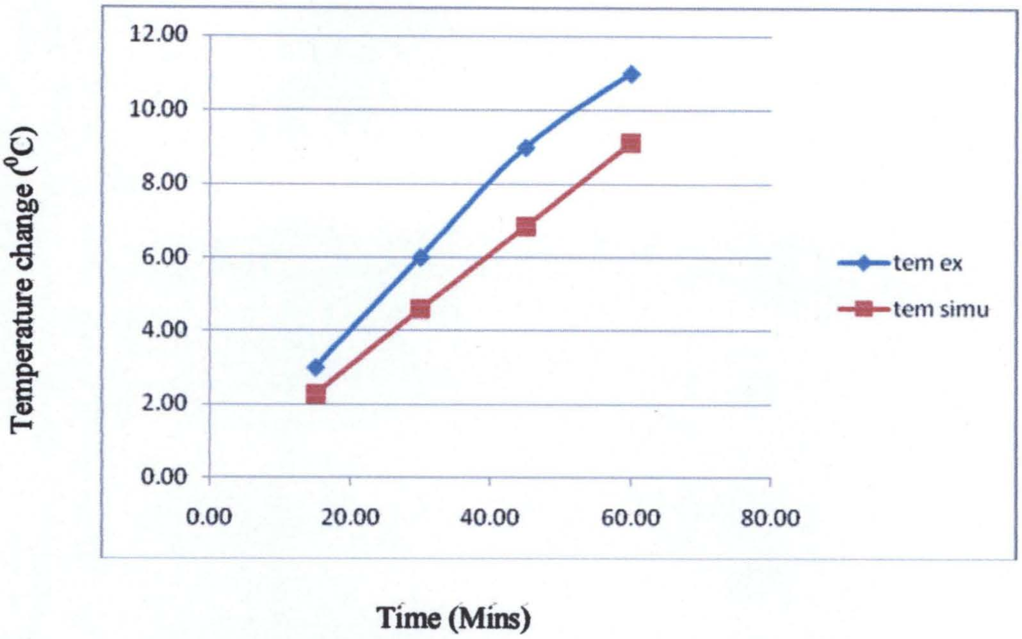
e = Subscript for Experiment

s = Subscript for Simulation

t = time (mins)

Covered = covered shoe boxes

Uncovered = uncovered shoe boxes



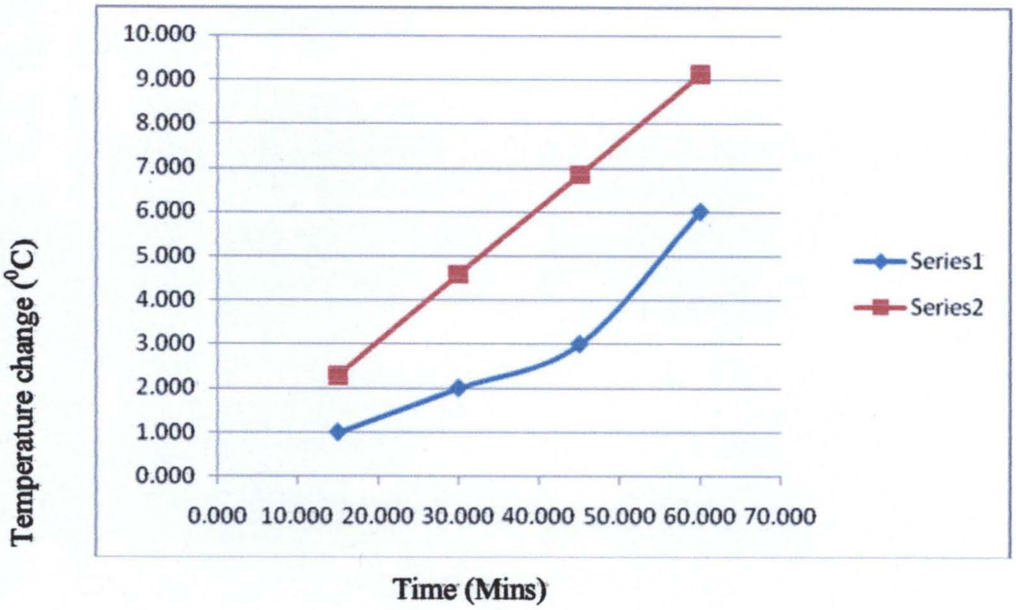
The Slope of Simulated Graph = 0.151733

The Slope of Experimental Graph = 0.18000

Figure 4.1: The Graph of Experimental and Simulated Temperature Change with Organic Soil

Table 4.2: Experimental and simulated temperature change with sharp sand soil

t/minutes	$T_{(covered)}$ °C	$T_{(uncovered)}$ °C	ΔT_e °C	ΔT_S °C	ΔT_{e-s} °C
15	37	36	1.000	2.279	-1.279
30	39	37	2.000	4.588	-2.588
45	42	39	3.000	6.837	-3.837
60	46	40	6.000	9.117	-3.117



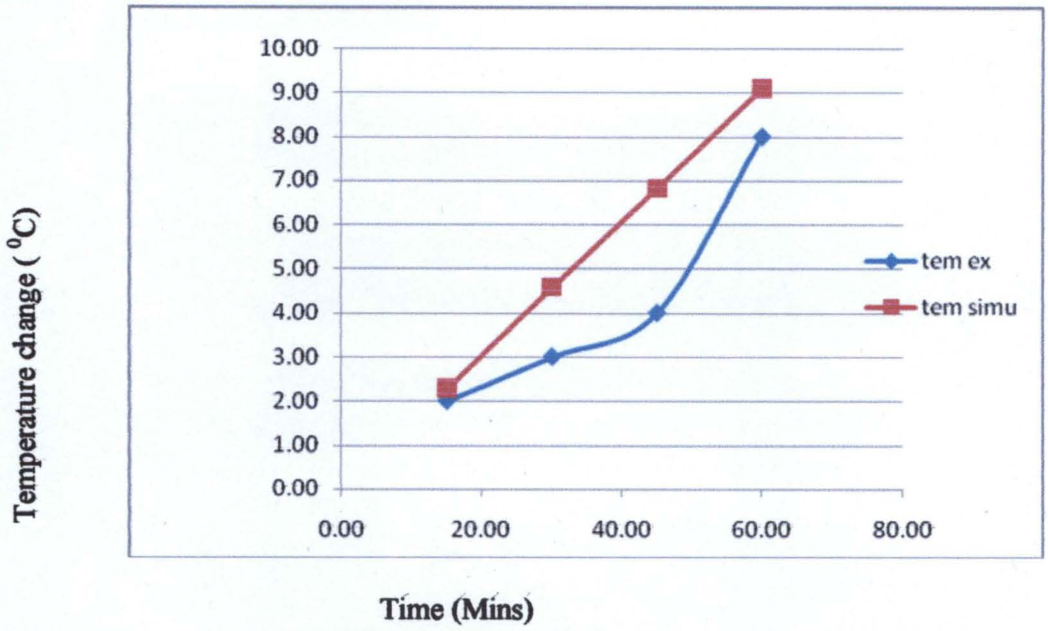
The Slope of Experimental Graph = 0.151733

The Slope of of Simulated Graph = 0.106667

Figure 4.2: The Graph of Experimental and Simulated Temperature Change with sandy Soil

Table 4.3: Experiment and simulated temperature change with chippings

t/minutes	$T_{(covered)}$ °C	$T_{(uncovered)}$ °C	ΔT_e °C	ΔT_S °C	ΔT_{e-s} °C
15	40	38	2.000	2.279	- 0.279
30	41	39	3.000	4.588	-1.588
45	44	40	4.000	6.837	-2.837
60	49	41	8.000	9.117	-1.117



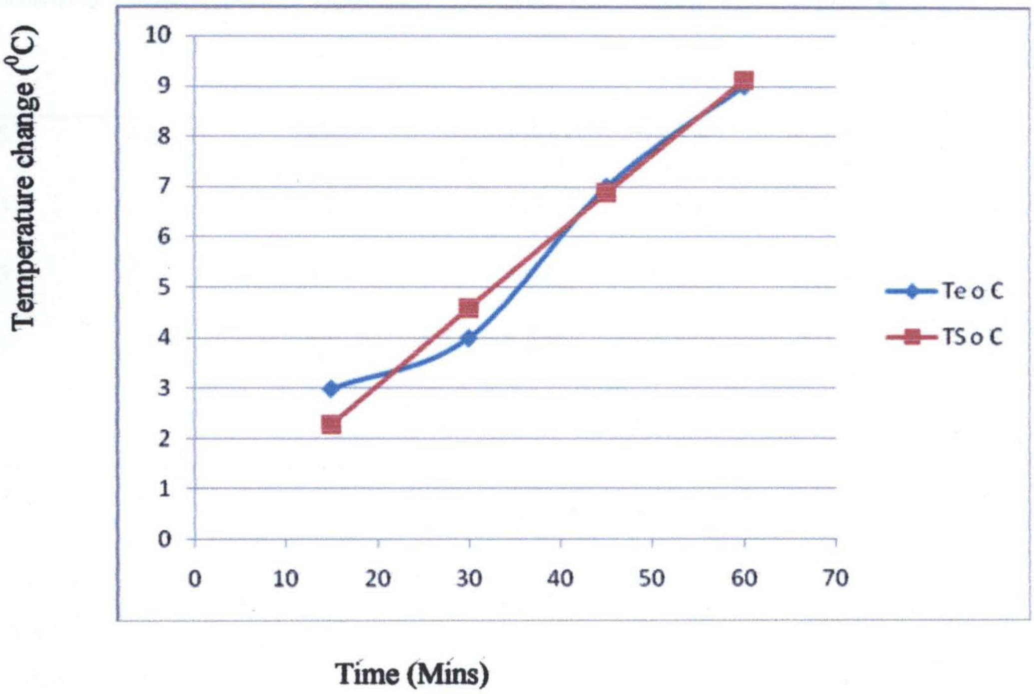
The Slope of Simulated Graph = 0.151733

The Slope of Experimental Graph = 0.126667

Figure 4.3: The Graph of Experimental and Simulated Temperature Change with Chippings

Table 4.4: Experiment and Simulated Temperature Change with Laterite

t/minutes	**$T_{(covered)}$ °C**	**$T_{(uncovered)}$ °C**	**ΔT_e °C**	**ΔT_S °C**	**ΔT_{e-s} °C**
15	41	39	2.000	2.279	-0.279
30	46	42	4.000	4.588	-0.588
45	49	43	6.000	6.837	-0.837
60	54	45	9.000	9.117	-0.117



The Slope of Experimental Graph = 0.153333

The Slope of Simulated Graph = 0.151753

Figure 4.4: The Graph of Experimental and Simulated Temperature Change with Laterite

5.2 Recommendations

The following are suggested for the future work:

1. Other experimental procedure other than this can be used to demonstrate the green house effect..
2. The research should be done to know the particular gas that conducts heat more among the green house gases and it should be recommended to minimize its emission..
3. Further work should be done to develop a model which will accommodate both quantity of gases in the atmosphere and it ability to transfer heat.

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APPENDIX A

CALCULATION OF THE MODEL PARAMETERS

$h = 56.78j/m^2(\mathrm{sec})o_C$

$When\,\mathrm{sec} = \frac{1}{60}min$

$(1)\;\; h = \frac{56j}{m^{2(\frac{1}{60}min)oC}} = \frac{56.78j}{m^2(0.01667)mino_C} = 3406.12j/m^2mino_C$

$\delta = 5.6696\ X\ 10^{-8}W/m^2K^4$ (Sinnoth, 1994)

$when\ k = \left(\frac{1}{273}\right),\ and\ sec\ \frac{1}{60}min$

$$\delta = \frac{5.6696\ x\ 10^{-8}j}{m^2\left(\frac{1}{273}\right)^4\left(\frac{1}{60}\right)min.oc}$$

$$\delta = \frac{5.6696\ x\ 10^{-8}j}{0.01667min.m^2\ x\ 1.8\ x10^{-10}o_C^4} = 18894.88j/m^2mino_C^4$$

$C_p - 371.58Jkg$ (Perry and chiton, 1984)

$\rho = 1.2928\frac{kg}{m^3}$ $(David, (2002)$ $\varepsilon = 0.612$ $(Sokolov, 2009)$

APPENDIX B

SIMULATION OF EQUATION 3.17 WITH MATLAB

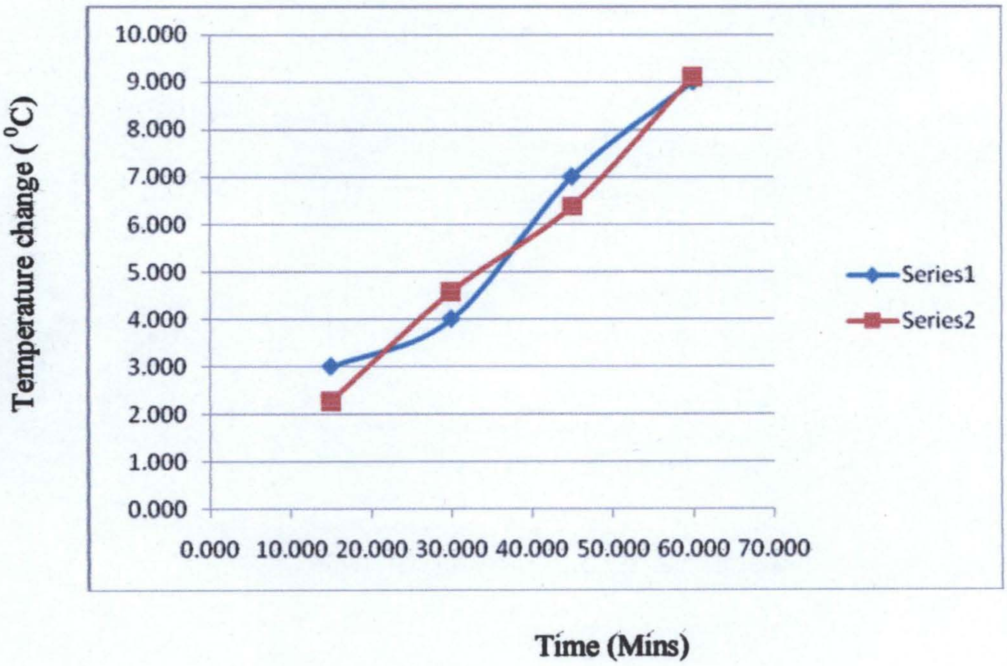
The definition of the letters used to simulate the equation are :

y = ΔT t = Δt
d = C_p v = ε
q = ρ e = δ

```
h=406.12;
v=18894.88;
q=371.58;
d=1.2928;
t=15;
T=15;
e=0.612;
y = ((0.009 * h * T + 0.000000006 * 4 * d * e * v * T ^ 4))/(d * q )* t ;
h=406.12;
v=18894.88;
q=371.58;
d=1.2928;
t=30;
T=15;
e=0.612;
y = ((0.009 * h * T + 0.000000006 * 4 * d * e * v * T ^ 4))/(d * q )* t ;
h=406.12;
v=18894.88;
q=371.58;
d=1.2928;
t=45;
T=15;
e=0.612;
y = ((0.009 * h * T + 0.000000006 * 4 * d * e * v * T ^ 4))/(d * q )* t ;
h=406.12;
v=18894.88;
q=371.58;
d=1.2928;
t=60;
T=15;
e=0.612;
y = ((0.009 * h * T + 0.000000006 * 4 * d * e * v * T ^ 4))/(d * q )* t ;
```

Table 4.5 Experiment and simulated temperature change with charcoal

t/minutes	$T_{(covered)}$ °C	$T_{(uncovered)}$ °C	ΔT_e °C	ΔT_S °C	ΔT_{e-s} °C
15	44	41	3.000	2.279	0.721
30	47	43	4.000	4.588	-0.588
45	53	46	7.000	6.837	0.163
60	57	48	9.000	9.117	-0.117



The Slope of Experimental Graph = 0.140000

The Slope of Simulated Graph = 0.14875

Figure 4.5: The Graph of Experimental and Simulated Temperature Change with Charcoal

4.1 Discussion of Results

This research work modeled the temperature change which arises in the atmosphere due to green house effect produced by various materials.

Table 4.1 represents the relationship between the temperature changes with respect to time at an interval of fifteen (15) minutes when the shoe box was filled with organic soil. The experimental and simulated results followed the same trend as the temperature changes increases with respect to time from zero (0) to sixty (60) minutes. This was due to the fact that green house gases trapped underneath the plastic sheet and soil occupied within the shoe boxes absorbed heat. The soil also re- radiates heat which was also absorbed by the green house gases. The two heats absorbed by green house gases and the soil caused the temperature of the thermometer to rise as shown. The effects of temperature change of the soil is primarily influenced by soil albedo, sensible heat flux, latent heat flux and heat flux of the soil as suggested by Rosenberg; (1974). The temperature change of the experiment (ΔT_e), in the time intervals of zero (0) to sixty (60) minutes is almost uniform implies that the heat transfer ability of organic soil is uniform through out the time intervals. The experimental result is a bit higher than that of simulated result. The difference between experimental and simulated results is because of the ideal conditions such as density, specific heat capacity, emissivity and Boltman Stefan Constant assumed during simulation which are real during the experiment. This difference in the values of Table 4.1 also results to a small difference in the rate of change of temperature with respect to time (Slope) of Figure 4.1 The slope of experimental result is lower than that of the simulated one and the simulated one showed a linear relationship with the experimental one in the Table 4.1 of both experimental and simulated results. It is also observed that the difference between the minimum and the maximum temperature is 0.721 and 1.883 respectively. In view of the above, Soil temperature fluctuates annually and daily affected mainly by variations in air temperature and solar radiation. The annual variation of daily average soil temperature at different depths can be estimated using a sinusoidal function (Hillel, 1982; Marshall and Holmes, 1988; Wu and Nofziger, 1999) .

Table 4.2 showed that the temperature change for simulated and experimental result when sharp sand was used. These temperature changes of experimental results are lower than when organic soil was used. The low temperature changes is due to the low soil albedo, low sensible heat flux, low soil heat flux, and low latent heat flux of sharp sand compared to that of organic soil. The properties that results to low heat of radiation is coupled with the presence of gases; hence heat absorbed by the sharp sand is lower than that of organic soil. According to Bristow *et al,* (2001) soil thermal properties are strongly influenced by the soil volumetric water content, volume fraction of solids and volume fraction of air. Air is a poor thermal conductor and reduces the effectiveness of the solid and liquid phases to conduct heat. The porosity of the sandy soil gave space to air in the sandy soil which results to low maximum temperature change of Table 4.2 and the difference between simulated and experimental temperature change all shown negative integers describing the temperature change of sandy soil. The result is also reported by Wiley (2006). Figure 4.2 shows that the slope of experimental temperature change is higher than that of the simulated one but they all followed the same trend and they display linear relationship.

The experimental results of Table 4.3 followed the same trend as in Table 4.2 and Table 4.1 but the temperature changes of this experimental result is higher than that of the sharp sand and lower than that of the organic soil caused by the differences in their properties such as soil albedo, sensible heat flux, latent heat flux and heat flux of the soil as suggested by Penuriasous, (2008). The experimental result showed a sharp rise in temperature change from 4 ° C to 8 ° C at time intervals of forty (45) minutes – sixty (60) minutes signifying a high heat flux of chippings at that temperature change (Rosenberg, 1974). The experimental and simulated results showed a linear relationship and the two results increased in temperature changes with respect to time from zero (0) to sixty (60) minutes. The Figure 4.3 shows that this experimental slope of chippings is higher than when sharp sand was used and lower than when organic soil was used which correlates with the results of the experiment and simulation.

The results of Table 4.4 also followed the same trend as in Table 4.1, Table 4.2 and Table 4.3 respectively with increase in temperature changes at interval of zero (0) to sixty 60 minutes. The experimental temperature changes are close to the simulated temperature changes at the same time interval while both the two temperature changes showing linear relationship with respect to time. The experimental temperature changes of this laterite is higher than that of the sharp sand and chippings but lower than that of the organic soil resulting from the amount of green house gases Wiley,(2006) and heat flux of the soils at those times as reported by Penuriasous, (2008). Figure 4.4 also showed that the slope of experimental temperature change is low and close to the simulated one. The slope of laterite is higher than that of the sharp sand and chippings but lower than that of the organic soil which obeys the same trend as in the table of results.

The results of Table 4.5 also followed the same trend as in Table 4.1, Table 4.2, Table 4.3 and Table 4.4 respectively. In this case, the temperature change in the experimental result of charcoal is higher than when sharp sand and chippings were used and is similar to when laterite was used but lower than when organic soil was used signifying a high heat flux according to Penuriasous, (2008) and the amount of green house gases. The temperature changes of this experimental result increased with respect to time. Figure 4.5 also showed that the slope of experimental temperature change is lower than the simulated one. The slope of this experiment is lower than that of the laterite, organic soil, and higher than that of the sharp sand and chippings.

The temperature changes all depend on properties of the soil such as soil albedo, sensible heat flux, heat flux and the latent heat flux of the materials reported by Penuriasous, (2008). and the absorption of heat and infrared rays by green house gases and transform them to thermal energy which results to increase in temperature of thermometer reported by Penuriasous, (2008).The transparent leather wrap also increases heating properties of the soil, the amplitude of the soil temperature and promote large radiation at the soil surface as a consequences the minimum and maximum temperature are increased reported by Streck, (1994)

The average slope of the temperature changes of experiment and simulation are 0.1413334 and 0.151153 with the percentage error of 0.98% which is negligible.

CHAPTER FIVE

5.0 CONCLUSIONS AND RECOMMENDATIONS

5.1 Conclusion

The following conclusions can be drawn from this work

1. The energy from the sun to the top surface of the shoe boxes and soil radiation have effects on temperature change when various soils were used for experiments. However charcoal gives the highest minimum experimental temperature change. This could be attributed to high sensitive heat flux of the charcoal that influences it to radiate heat. This clearly demonstrates that the green house effect also depends on the top surface of the soil; which gives rise to it.
2. The results also showed that there is a linear relationship between the temperature changes in the experimental and simulated results as shown in the Figures 4.1 to 4.5 respectively, but little deviation from simulated results because of the ideal conditions assumed during simulation. However the temperature changes vary with the types of materials used for the experiment.
3. The proposed deterministic model developed can be used for predicting temperature changes of the materials used for the experiment to demonstrate the green house effect.
4. This model is able to evaluate the rate of change of temperature with respect to time.(Slope) and the percentage error between the highest and the lowest slope is 0.98%.